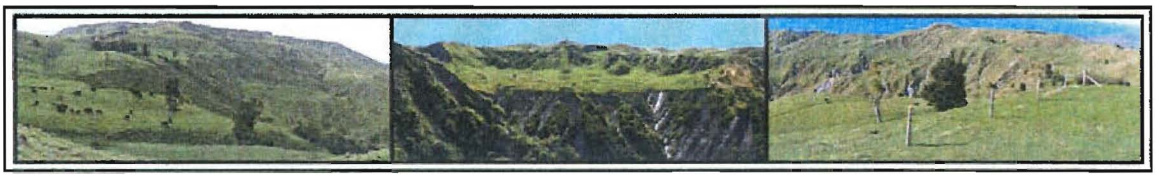
Rock mass defect controlled deep-seated landslides in Tertiary soft rock terrain: Implications for landscape evolution



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Abstract

Rock mass defect controlled deep-seated landslides are widespread within the deeply incised landscapes formed in Tertiary soft rock terrain in New Zealand. The basal failure surfaces of deep-seated slope failures are defined by thin, comparatively weak and laterally continuous bedding parallel layers termed *critical stratigraphic horizons.* These horizons have a sedimentary origin and have typically experienced some prior tectonically induced shear displacement at the time of slope failure. The key controls on the occurrence and form of deep-seated landslides are considered in terms of rock mass defect properties and tectonic and climatic forcing.

The selection of two representative catchments (in southern Hawke's Bay and North Canterbury) affected by tectonic and climatic forcing has shown that the spatial and temporal initiation of deep-seated bedrock landslides in New Zealand Tertiary soft rock terrain is a predictable rather than a stochastic process; and that deep-seated landslides as a mass wasting process have a controlling role in landscape evolution in many catchments formed in Tertiary soft rock terrain.

The Ella Landslide in North Canterbury is a deep-seated (~85 m) translational block slide that has failed on a 5 – 10 mm thick, kaolinite-rich, pre-sheared critical stratigraphic horizon. The residual strength of this sedimentary horizon, (C'_R = 2.6 – 2.7 kPa, and Θ'_R = 16 – 21°), compared to the peak strength of the dominant lithology (C' = 176 kPa, and Θ' = 37°) defines a high strength contrast in the succession, and therefore a critical location for the basal failure surface of deep-seated slope failures. The (early to mid Holocene) Ella Landslide debris formed a large landslide dam in the Kate Stream catchment and this has significantly retarded rates of mass wasting in the middle catchment. Numerical stability analysis shows that this slope failure would have most likely required the influence of earthquake induced strong ground motion and the event is tentatively correlated to a Holocene event on the Omihi Fault. The influence of this slope failure is likely to affect the geomorphic development of the catchment on a scale of $10^4 – 10^5$ years.

In deeply incised catchments at the southeastern margin of the Maraetotara Plateau, southern Hawke's Bay, numerous widespread deep-seated landslides have basal failure surfaces defined by critical stratigraphic horizons in the form of thin (< 20 mm) tuffaceous beds in the Makara Formation flysch (alternating sandstone and mudstone units). The geometry of deep-seated slope failures is controlled by these regularly spaced (~70 m), very weak critical stratigraphic horizons (C'_R = 3.8 – 14.2 kPa, and Θ'_R = 2 – 5°), and regularly spaced (~45 m) and steeply dipping (~50°) critical conjugate joint/fault sets, which act as slide block release surfaces. Numerical stability analysis and historical precedent show that the temporal initiation of deep-seated landslides is directly controlled by short term tectonic forcing in the form of periodic large magnitude earthquakes. Published seismic hazard data shows the recurrence interval of earthquakes producing strong ground motions of 0.35g at the study site is every 150 yrs, however, if subduction thrust events are considered the level of strong ground motion may be much higher.

Multiple occurrences of deep-seated slope failure are correlated to failure on the same critical stratigraphic horizon, in some cases in three adjacent catchments. Failure on multiple critical stratigraphic horizons leads to the development of a "stepped" landscape morphology. This slope form will be maintained during successive accelerated stream incision events (controlled by long term tectonic and climatic forcing) for as long as catchments are developing in this specific succession. Rock mass defect controlled deep-

seated landslides are controlling catchment head progression, landscape evolution and hillslope morphology in the Hawke's Bay study area and this has significant implications for the development of numerical landscape evolution models of landscapes formed in similar strata. Whereas the only known numerical model to consider deep seated landslides as an erosion process (ZSCAPE) considers them as stochastic in time and space, this study shows that this could not be applied to a landscape where the widespread spatial occurrence of deep-seated landslides is controlled by rock mass defects.

In both of the study areas for this project, and by implication in many catchments in Tertiary soft rock terrain, deep-seated landslides controlled by rock mass defect strength, spacing and orientation, and tectonic and climatic forcing have an underlying control on landscape evolution. This study quantifies parameters for the development of numerical landscape evolution models that would assess the role of specific parameters, such as uplift rates, incision rates and earthquake recurrence in catchment evolution in Tertiary soft rock terrain.

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Chapter One

1.0 Introduction

Landscapes formed in stratified Tertiary soft rock successions are often subject to large volume, deep-seated slope instability. Slopes may fail on basal shear surfaces dipping at just a few degrees, and the sensitivity of such failures to the contribution of various factors is often poorly defined and understood. The failure surfaces of many deep-seated landslides in Tertiary soft rock are coincident with stratigraphic orientation, and very low strength stratigraphic horizons may be a crucial component which has a controlling influence on this form of slope instability.

Questions frequently arise regarding the triggering mechanism for such large slope failures and a common consideration is earthquake induced strong ground motion. Earthquakes are very well documented as triggering historical landslides (e.g. Keefer, 1984), however, determining trigger mechanisms for prehistoric landslides is more difficult. While it is often inferred that prehistoric landslides might be earthquake triggered (e.g. Pettinga, 1980; Adams, 1981; Read et al., 1992; Beetham, 1994) it is generally difficult to demonstrate that this is so. It is important in the context of landscape evolution, sediment production and hazard assessment that the relative influence of strong ground motion on slope stability is well understood.

Sediment production from catchments developing in Tertiary soft rock terrain is sometimes considered to be dominated by shallow regolith landslides (e.g. Crozier et al., 1992), and this could be interpreted as considering that shallow landslides dominate mass movement in such landscapes, controlling denudation and hillslope development. The specific focus on the high occurrence of shallow mass movement processes reflects the massive increase in sediment production subsequent to extensive land clearance following European settlement. This trend is well represented in east coast North Island soft rock areas, with sediment production rates up to 17 times those under native forest cover (e.g. Page and Trustrum, 1997; Wilmshurst, 1997). In some such landscapes, however, periodically active deep-seated bedrock landslides cover a significant portion of the land surface and therefore must be considered to have an influence on both sediment production and landscape development. The topographic form of a landscape will vary depending on the mode of erosion processes acting on it. It may be generalised that in a shallow mass movement dominated landscape the hillslope form will tend toward dendritic drainage patterns with higher angle slopes, while in a landscape containing a significant proportion of deep-seated mass movement the hillslope form is more likely to be characterised by relatively subdued (lower angle) slopes dissected by deeply incised drainage networks.

When a significant portion of any selected landscape is affected by numerous deep-seated mass movements then it must be assumed that these are playing an underlying, if not controlling, role in the evolution of that landscape. To quantify the role of deep-seated landslides in landscape evolution it is necessary to obtain data defining controls on the geometry and spatial and temporal initiation of deep-seated slope failures, and with knowledge of such controls the occurrence of deep-seated landslides may be considered a predictable process.

1.1 Thesis objectives

The overall objective of this thesis is to define, and where possible quantify, controls on the geometry and spatial and temporal initiation of deep-seated bedding controlled landslides in Tertiary soft rock terrain and to consider these in the context of landscape evolution.

The main objectives of this project are:

- To select field sites that are representative of a larger population of catchments developing in Tertiary soft rock successions, in tectonically active areas of New Zealand, that are affected by deep-seated slope failures
- To identify and investigate the development of critical stratigraphic horizons in Tertiary soft rock successions which act as failure surfaces for deep-seated landslides
- To quantify parameters required for numerical slope stability modelling, including critical rock mass defect properties such as strength and geometry
- To back analyse selected prehistoric deep-seated soft rock landslides to quantitatively assess the role of strong ground motion as a triggering mechanism; and
- To consider the evolution of selected landscapes, in terms of the role of deep-seated landslides and tectonic and climatic forcing, with a view to providing parameters for the development of numerical landscape evolution models.

While the intention of this research is to outline quantified controls on deep-seated landslides that might form the basis of numerical landscape evolution models, the actual development of such models is beyond the scope of the project.

1.2 Slope stability factors in soft rock terrain

1.2.1 Distribution and definition of Tertiary soft rock in New Zealand

Internationally there is no standardised definition or terminology for these materials (Hawkins, 2000), which are variously described as "hard soils", "indurated soils", "weak rocks", "weak rock materials", "low strength rocks" and more (e.g. Oliveira, 1993). In North America and other parts of the world similar materials are frequently described as shales (e.g. Wilkenshaw and Santi, 1996). In New Zealand these soft rock successions are widely distributed (Figure 1.1) and primarily consist of overconsolidated marine sequences of limestone, sandstone, siltstone and mudstone.

Definition and rock material characteristics

Several authors have discussed New Zealand sedimentary rocks which are loosely classed as soft rocks (e.g. Brown, 1974; Borrie et al., 1980; Read et al., 1981; Bell and Pettinga, 1984, 1988; Pettinga and Bell, 1992; Prebble, 1992), and these can be generally defined as Upper Cretaceous, Tertiary and Quaternary sedimentary rocks with two primary characteristics:

- They have low strength, typically the unconfined compressive strength would be from one to several MPa and less than ten MPa; and

- They are unstable in water. For coarser grained lithologies (sandstones) this leads to particle disaggregation upon immersion in water due to weak particle bonds, while for finer grained lithologies (mudstones and siltstones) these are prone to slake degradation related to: i) wetting and drying cycles acting on shrink swell prone clay minerals; and, ii) a pore pressure differential upon immersion in water, forcing the (porous) rock material apart.

Soft rock sequences are typically overconsolidated (weakly lithified) and material strength properties often straddle the boundary between engineering soil and rock (e.g. Wilkenshaw and Santi, 1996). Soft rock successions commonly contain inherited stresses which can cause exfoliation slabs to develop parallel to exposed faces (e.g. Thompson, 1981; Huppert, 1988; Prebble, 1992).

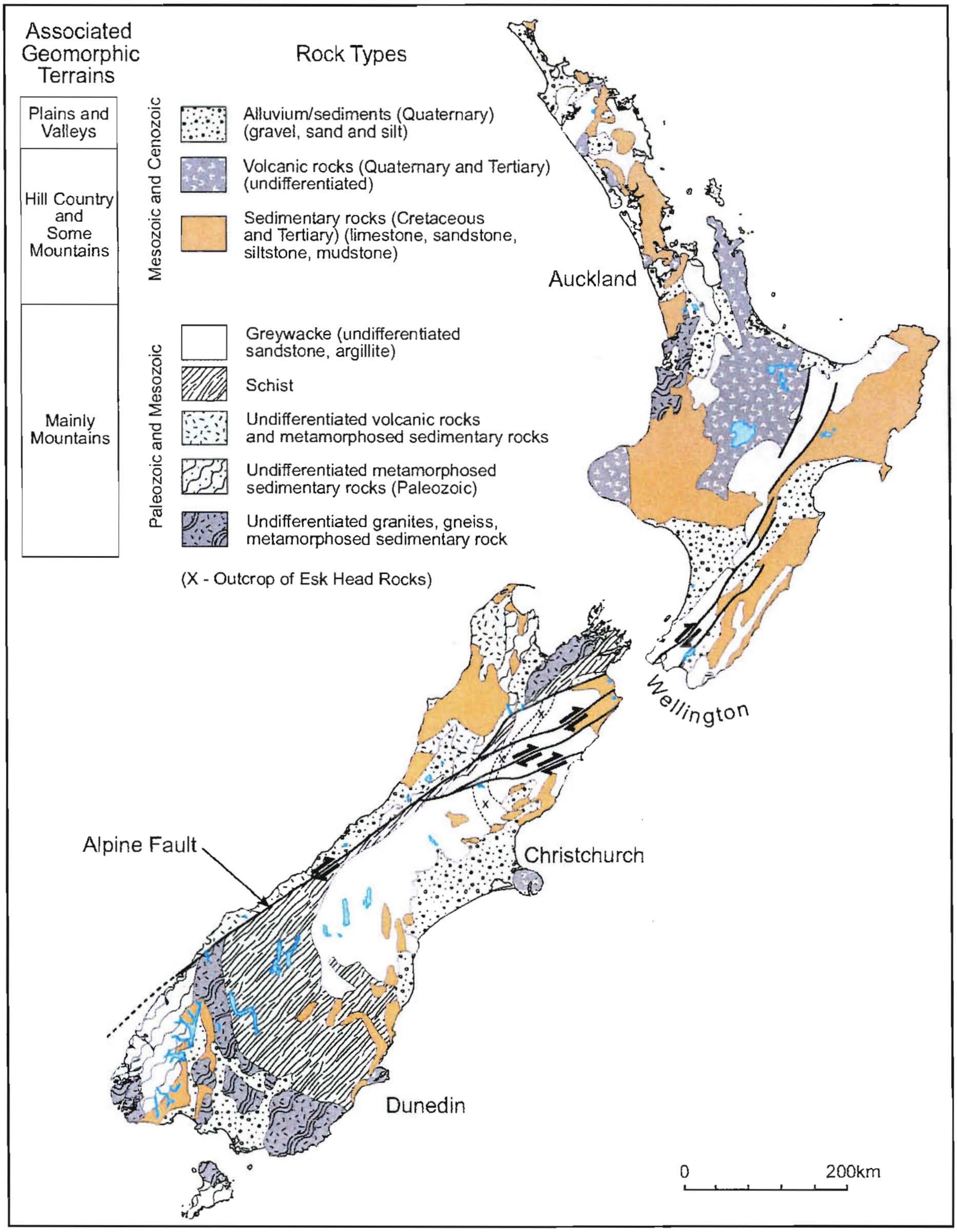


Figure 1.1: Distribution of rock types within New Zealand. Soft rocks are primarily represented by the Cretaceous – Tertiary sedimentary successions (shown in brown). Figure modified from Pettinga (2001).

Rock mass characteristics

One of the most critical parameters controlling bedrock slope stability relates to rock mass defect characteristics (Bell and Pettinga, 1988). Comparatively weak bedding parallel defects can occur at bedding contacts in clearly bedded sequences, and as bedding partings in sequences where bedding is otherwise not clearly defined (J. Pettinga pers. comm. 2005). As well as bedding contact related defects, depositional layers such as volcanic ash horizons with contrasting strength, grainsize and cementation properties to the stratigraphically adjacent materials may also define weak horizons within the stratigraphy.

Defects in the form of jointing and faulting can provide discrete rock mass boundaries, defining the geometry of blocks that may be susceptible to failure. These defects are likely to have a lower tensile and/or shear strength than the overall rock mass, and so are more likely to form release surfaces for slope failures than intact rock material.

1.2.2 Deep-seated slope instability in New Zealand soft rock terrain

Slope instability can be considered to be ubiquitous in Tertiary soft rock terrain (Bell and Pettinga, 1988), and many deep-seated landslides are documented within such terrain (e.g. Brown, 1974; Stout, 1977; Coombs and Norris, 1981; Smale et al., 1982; Bell and Pettinga, 1988; Pettinga, 1992; Prebble, 1992). The primary form of the initial movement of a deep-seated soft rock slope failure is likely to be as a translational block or wedge slides, although initially intact slide blocks may rapidly break up and degrade into debris slides or flows (terminology based on Cruden and Varnes, 1996).

Bedding parallel failure surfaces as "critical stratigraphic horizons"

The initial failure of deep-seated landslides in stratified sedimentary sequences is typically on bedding parallel failure surfaces of low shear strength. As these surfaces define where landslides occur stratigraphically, and are likely to have a controlling influence on slope stability, they might be thought of as *critical stratigraphic horizons*. Stout (1977) was one of the first authors to document the importance of stratigraphically defined failure surfaces in New Zealand when he recognised that the aerially extensive (~18 ha) Utiku Landslide had failed on a single bedding surface in central North Island soft rock. The relevant stratigraphic sequence in the area was recognised to contain montmorillonite layers just a few millimetres thick, which have allowed very large blocks to slide parallel to bedding.

Large landslides failing on very thin shear surfaces are now well documented in New Zealand soft rock terrain (e.g. Coombs and Norris, 1981; Pettinga, 1987a; Bell and Pettinga, 1988; Fell et al., 1988; Pettinga and Bell, 1992; Prebble, 1992), and many studies have considered the importance and development of bedding parallel surfaces as basal shears for slope failures (e.g. Skempton, 1964, 1966; Bjerrum, 1967; Sugden et al., 1977; Pinckney et al., 1979; Barton, 1984, 1988; Fell et al., 1988; Hutchinson and Anonymous, 1995; Hart, 2000; Hamel and Hart, 2001).

A commonly observed feature of stratigraphic horizons that define the basal shear surface of a landslide is the occurrence of shear displacement prior to slope failure (e.g. Pettinga, 1987a; Fell et al., 1988; Wang et al., 2003), and this may be considered as "pre-shearing". When a layer within a soft rock sequence is pre-sheared, the strength of that layer is likely to be at or near its residual strength and the layer hence provides a very weak horizon within the rock mass. The development of pre-shearing in rock masses has received much attention in terms of both folded sequences (e.g. Tanner, 1989) and in horizontal or near-horizontal sequences, such as the review by Hart (2000). The large number of mechanisms proposed for pre-shear development precludes consideration of them all, and only those relevant to this study are included here.

Flexural slip
Shear development in tilted or folded strata has long been attributed to a tectonic origin (e.g. Skempton, 1966; Sugden et al., 1977) and a well documented mechanism for this is flexural slip (e.g. Tanner, 1989; Hutchinson and Anonymous, 1995). Sometimes termed "layer on layer slip" this occurs when folding of layered strata is accommodated by inter-bed movement (Figure 1.2), analogous to bending a pack of cards and allowing the cards to slide over one another and so enable the pack to fold or deform as a single slab. This causes frictional movement which can develop gouge material between beds, and typically defines a locus of strength contrast.

Progressive failure
Skempton (1964) considered how landslides in overconsolidated clays could be preceded by the development of a continuous sliding surface, and Bjerrum (1967) expanded on this to consider how overconsolidated materials can contain recoverable strain energy due to elastic deformation during consolidation. Differential stresses will occur in the direction of least resistance and if these stresses exceed the material or rock mass strength then a slip

surface will develop for as far as the stress exceeds the strength. Slope parallel release surfaces are common in such overconsolidated rocks (what may be termed stress release fractures or exfoliation defects), however, Bjerrum (1967) also sites cases where these slip surfaces progress in the direction of bedding.

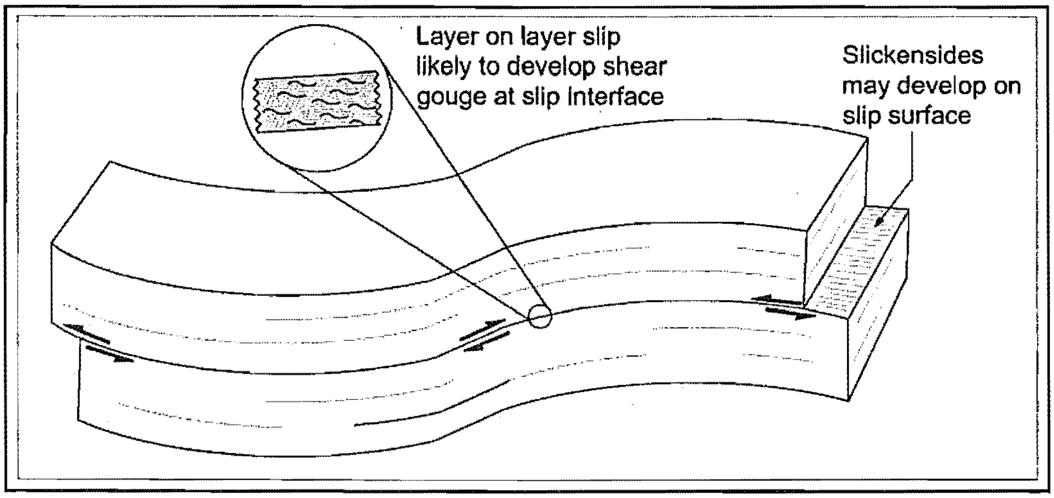


Figure 1.2: Schematic representation of the flexural (layer on layer) slip mechanism of a fold and subsequent shear development. Plastic deformation at bedding contacts facilitates folding of strata and may induce shear gouge development.

Figure 1.3 shows how progressive failure might occur in the walls of a deeply incised gully. In this situation the stratigraphic sequence contains thin weak layers, with a high strength contrast to surrounding strata, and it is likely that all movement will be concentrated on these (indicated by arrows in Figure 1.3). A shear gouge may also be developed (Figure 1.3 inset). Shear development may progress into the slope until the point where recoverable strain energy or horizontal stress (σ_H) is less than the strength of the layer on which slip is occurring. As this shear is preferentially occurring on pre-existing weak horizons, this could be thought of as preferential progressive failure.

It is apparent that whatever the mechanism for developing shear in such layers, there is likely to be a sedimentological reason for their location within the stratigraphy. Specific layers in bedded clay rich stratigraphy may facilitate failure due to inherent shear strength properties, rather than solely due to an adverse stress condition which would be expected to initiate failure in a truly homogeneous sequence. Although the actual shear surface may be developed by one of the mechanisms discussed above, it is likely to be exploiting a weak lithological layer or sedimentary characteristic. Possible sedimentary reasons for the location of such layers have been considered (e.g. Barton, 1988), and the most plausible causes are considered to be: i) a permeability contrast; ii) an increase in clay content; iii) a change in clay mineralogy; or, iv) a strength contrast such as increased cementation. All these sedimentary characteristics can provide a shear strength contrast or "defect" in the stratigraphy which may cause a particular horizon to shear in preference others. A stratigraphic defect that has no clearly defined material contrast either side (i.e. the sedimentary characteristics appear to be the same above and below) may be considered as a "bedding parting" which may simply be defined by a brief pause in sedimentation.

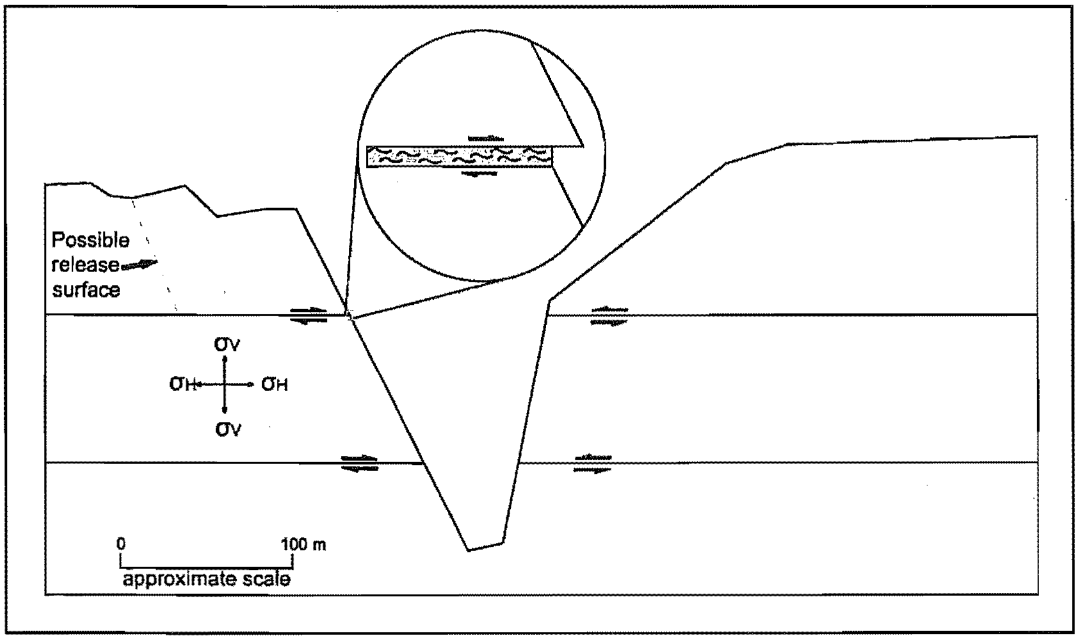


Figure 1.3: Schematic diagram showing how differential lateral rebound of a laterally overconsolidated sequence of near horizontal strata could cause pre-shear development.

1.3 Landscape forcing factors

The influence of tectonic and climatic processes on landscapes and the degradation processes acting on them is particularly significant (e.g. Bull, 1991; Burbank and Anderson, 2001). Tectonic and climatic processes in the context of landscape development can be considered at different time scales and may be termed "landscape forcing".

1.3.1 Long-term landscape forcing

Long-term "tectonic forcing" encompasses the tectonic influences on a landscape at the 10^4 – 10^6 yr scale. Processes referred to as tectonic forcing include:

- Uplift of bedrock mass: By raising parts of the landscape, relative to both sea-level and other parts of the landscape, the action of erosional processes is enhanced.

- Tilting of strata: This affects the way in which mass wasting processes are able to occur by introducing stratigraphic dips.

- Folding of strata: The geometry of folded sequences can have a significant influence on the topographic form of a landscape and valley and range systems are often coincident with fold belt orientations.

Of particular importance to landscape development is the changing elevation of the landscape with respect to sea-level. Long-term "climatic forcing" in the form of 10^4 yr orbitally forced glacial and interglacial cycles has a significant influence on global sea-levels (e.g. Bull, 1991; Burbank and Anderson, 2001). Long-term climate cycles will also affect landscapes by varying the size of glaciers and rivers and hence the amount of work they can do in modifying the landscape. The response of stream networks to sea-level variation is of significant importance to landscape development in terms of the concept of the "base level of erosion" (Bull, 1991). This defines the equilibrium stream profile below which the stream bed can not degrade and where neither erosion nor deposition is occurring. The reference level is generally taken to be sea-level, as a river or stream is unlikely to erode to an elevation below its final destination. This concept is of particular importance in coastal catchments where any variation in relative sea-level will rapidly disturb the base level of stream networks.

1.3.2 Short-term landscape forcing

In the short-term tectonic and climatic forcing provide the means to initiate mass movement processes. Short-term tectonic forcing refers to large magnitude earthquakes inducing strong ground motion in the landscape with $10^2 - 10^3$ yr return periods. Short-term climatic forcing refers to variation in climate on both the annual (storm event) and decadal (climate regime) scale.

There are well documented instances of landslide triggering by mechanisms of short-term climatic forcing (e.g. Crozier, 1986) and short-term tectonic forcing (e.g. Keefer, 1984). While climatic events are particularly important for triggering shallow mass movement, it is well documented that deep-seated landslides can occur co-seismically with large magnitude earthquakes, for example several large landslides triggered by the 1931 M_w 7.8 Hawke's Bay earthquake (Marshall, 1933) and the 180 million m^3 Bairaman Valley debris flow in Papua New Guinea triggered by an M_L 7.1 earthquake (King et al., 1989). Studies have also demonstrated the likelihood of a seismic trigger for pre-historic landslides (e.g. Jibson and Keefer, 1993; Crozier et al., 1995). For further information on mechanisms of slope destabilisation Wieczorek (1996) provides an overview of general landslide triggering mechanisms and Kramer (1996) provides detail on the nature of slope destabilization by earthquake ground motion.

It is clear that seismic activity plays an important role in triggering deep-seated landslides and is something which merits further investigation. The influence of climate and tectonics on slope stability is addressed further in Chapters 2 and 4, and specific discussion on earthquake triggering of landslides is included in Chapter 5 of this study.

1.4 Numerical modelling of landscape evolution

1.4.1 Landscape evolution – form and process

Much of the research on landscape evolution is concentrated on hillslopes. Hillslopes provide sediment, which in turn feeds to basins via rivers/floodplains, lakes and oceans. As such, they can be considered a starting point for the "sediment cycle", although it would be short-sighted to consider them in complete isolation. The same processes which remove sediment from hillslopes contribute greatly to the instability of them (e.g. rivers/streams remove debris from lower slopes and undercut slopes leading to further mass movement). Hillslopes may be viewed as a system of stores and transfers (Selby, 1982), as material is eroded, transported and deposited by a variety of processes. If the processes removing material from hillslopes are quantified, as well as the underlying controls on these processes and the landscapes response to them, it is possible to model how a particular landscape will evolve over time.

Davis (1909) is generally attributed as being the first to propose that the surface of the Earth's development is controlled by and subject to cycles. His proposals, however, are partially discounted as he failed to consider the role of continued tectonic uplift, coincident with landscape degradation. Schumm and Lichty (1965) brought to light the important aspect of time in relation to scale, highlighting how the distinction between cause and effect in landform development are a function of both the available time and the spatial extent of the landscape. Implicit in this is that any part of a geomorphic system should not be considered in either spatial of temporal isolation, but rather integrated within the overall framework of landscape development.

Computer based landscape evolution models appeared in the late 1970's (e.g. Anhert, 1976) and a dramatic advance in computer power in the last 20 years has allowed the development of computer based methods of analyzing hillslope development that more accurately represent fluvial and slope processes. Coulthard (2001) presents a review of selected landscape evolution modelling software. Computer based landscape evolution models can use mathematical representation of critical processes to assess the influence and sensitivity of interactions between mass movement, lithology, tectonics and climate on landscapes at hillslope, catchment and mountain range scale.

1.4.2 Landslides in landscape evolution models

Many numerical landscape evolution models consider slope development either by diffusive processes (e.g. Kooi and Beaumont, 1994) or processes which instantaneously lowers a slope above a defined threshold (e.g. Howard, 1994; Tucker and Slingerland, 1994). Bedrock landslides are incorporated into very few landscape evolution models, despite being an important process in the evolution of a variety of landscapes (Densmore et al., 1996; 1998). Roering et al. (in press) describe how a landscape developing in a marine sedimentary soft rock sequence can be shown to have significant topographic variation as defined by the influence of deep-seated landslides. Areas in which deep-seated landslides dominate show a more subdued topography compared to the steep and dissected terrain where stream incision and shallow landslide processes dominate. The relative occurrence of deep-seated landslides can be directly linked to bedrock structure and lithology.

The landscape evolution model ZSCAPE has been used to demonstrate the importance of bedrock landslides in the development of a normal fault bounded mountain range (Densmore et al., 1998; Ellis et al., 1999). ZCSAPE is a three dimensional model which simulates landscape evolution by tectonic and geomorphic processes in a finite difference grid. In terms of landslide process, the model defines where bedrock landslides occur, their size, and where the debris will be distributed. By considering the landslides to be stochastic the model ignores climatic and seismic variability as triggering mechanisms. The model also considers the medium to be essentially homogeneous, in that there is no allowance for defects or lithological heterogeneity. This limitation of the model makes it solely applicable to highly fractured rock masses, such as areas of Southern Alps Torlesse (indurated and highly fractured sandstones and argilites).

Many rock masses, such as the bedded Tertiary soft rock sequences, have distinct directionality that controls where and in what form deep-seated landslides will occur. If deep-seated landslides are to be considered in a model of landscape development in such a rock mass, then it will be necessary to parameterise controls on location, triggering, size and geometry of failures. Whereas deep-seated landslides are often considered to be a stochastic process (e.g. Densmore et al., 1998; Ellis et al., 1999) it is likely that once these parameters are understood and quantified, the spatial and temporal occurrence of deep-seated, rock mass controlled slope failures in soft rock terrain can be considered as less stochastic and more predictable.

Further consideration of landscape evolution modelling is addressed in Chapter 6, with specific reference to selected field sites for this project considered representative of large areas of New Zealand soft rock terrain. As mentioned in Section 1.1, the development of a numerical landscape evolution model is beyond the scope of this project. It is hoped that quantification of key controls on deep-seated landslides with respect to landscape development in a heterogeneous medium will allow them to be numerically predictable in both time and space.

1.5 Choice of representative sites within the New Zealand setting

In this project the objectives (as stated in Section 1.1) require specific field sites that can be considered representative of the wider areas of New Zealand soft rock terrain (refer Figure 1.1) that are affected by deep-seated landslides. Two areas have been selected, in coastal Southern Hawke's Bay and coastal North Canterbury, where deep-seated landslides occur and have potentially been triggered by seismic activity.

Coastal soft rock catchments in Hawke's Bay have received attention in the past (e.g. Pettinga, 1980; Leith, 2003; Mackey, 2003), and the evolution of these deeply dissected catchments is clearly being influenced by the presence of numerous instances of deep-seated slope instability. In North Canterbury coastal soft rock catchments are also known to be effected by large scale bedrock slope failures (e.g. Smale et al., 1982; Justice, 1994; Geotech Consulting Ltd, 2002). In comparison to Hawke's Bay the North Canterbury catchments contain few instances of large landslides, however, these are inferred to have a significant effect on catchment evolution.

The two chosen field sites (in Southern Hawke's Bay and North Canterbury) represent landscapes where two different situations are recognised; in Southern Hawke's Bay numerous widespread deep-seated slope failures are dominating the rate and form of landscape development, while in North Canterbury spatially and temporally infrequent deep-seated slope failures perturb catchment development and significantly affect the form of the landscape and processes acting within catchments.

1.6 Thesis organisation

This thesis includes seven chapters which logically progress toward a final conclusion by detailed analysis of the objectives stated in Section 1.1.

Chapters 1 & 2 give a general background to the context and methods applied in the thesis. They are both primarily literature reviews, with Chapter 1 focusing on general background information pertinent to this study. Chapter 2 provides a detailed overview of relevant regional to catchment scale geology and geomorphology for both study sites.

Chapter 3 outlines methods used for field data and sample collection, sample preparation and testing, and provides results from field investigations and geotechnical testing.

Chapters 4 & 5 discuss how the collected data was analysed to both develop models of the chosen landslides and to assess their stability.

Chapter 6 focuses on the role that these large landslides play in landscape evolution, and how they might be incorporated into numerical landscape evolution models.

Chapter 7 concludes the thesis with a discussion of the study findings and considers the relevance of these to the numerical modelling of landscape evolution.

1.7 Summary

This chapter has set out research objectives and provides a context for this study based around an initial literature review. The contextual information specifically focuses on factors which have a controlling influence on deep-seated slope stability, and how deep-seated slope failure effects landscape development.

New Zealand contains extensive areas of Upper-Cretaceous and Tertiary soft rock terrain. These soft rock materials are weak and commonly contain stratigraphically controlled horizons at residual strength which allow large-scale slope failures to occur by utilizing these critical horizons as failure surfaces. Deep incision, due to base level lowering in response to tectonic and climatic factors in Quaternary times, has allowed these horizons to daylight in the landscape. Worldwide there are many documented cases of deep-seated landslides being triggered by earthquake activity and this may also play a significant role in the widespread occurrence of deep-seated landslides in New Zealand soft rock terrain. The only numerical landscape evolution model known to consider deep-seated landslides (ZSCAPE) considers them a stochastic process occurring in a homogeneous medium.

The hypothesis which forms the basis for this research project is that in bedded soft rock sequences in New Zealand the initial movement of deep-seated landslides is controlled by material heterogeneity, and if the spatial and temporal controls on the form and initiation of these landslides can be quantified, deep-seated landslides may be considered predictable (as opposed to stochastic) over geomorphic time scales.

Chapter Two

2.0 Geological and geomorphological setting

2.1 Introduction

This chapter provides an overview of the geological and geomorphological setting for the selected study sites in Southern Hawke's Bay and North Canterbury, based on a literature review of previous publications at both a regional and more detailed (local) scale. The purpose of this chapter is to provide the context in which deep-seated landslides are occurring within landscapes in Tertiary soft rock terrain. More detailed site specific data collected as part of field investigations for this project are presented in Chapter 3.

2.2 New Zealand tectonic setting

The New Zealand continent straddles the active plate boundary zone of the Pacific and Australian Plates (Figure 2.1). To the north plate convergence > 40 mm/yr is reflected in subduction of the Pacific plate under the Australian plate from the northern South Island to the Hikurangi Trough and the Kermadec Trench (Walcott, 1978). To the south slower (~30 mm/yr) plate convergence is significantly more oblique and the Australian plate is subducted under the Pacific Plate. The New Zealand landmass straddles these two opposite dipping subduction zones (west dipping in the north and east dipping in the south) and oblique plate convergence and collision is reflected in the 3 km+ of topographic uplift in the Southern Alps. Significant structural features of the plate boundary include the Hikurangi Trough, Marlborough Fault Zone and the Alpine Fault, and the New Zealand landmass can be divided into different tectonic provinces based on upper crustal fault behaviour (e.g. Berryman and Beanland, 1991). The two provinces of specific interest to this project are the zone of thrust faults of the highest emergent accretionary ridge, between the Hikurangi Trough and the North Island Shear Belt, and the zone of thrust faults (North Canterbury Fold and Fault Belt of Pettinga et al., 2001) to the south east of the Marlborough fault zone (refer Figure 2.1).

The Hikurangi Margin, on the east coast of the North Island, is an imbricate frontal accretionary wedge forming in response to oblique subduction (Lewis and Pettinga, 1993; Barnes et al., 2002). In the overriding Australian Plate this is represented as a zone of thrust faulting and thrust propagated folding, and emergent ridges of the accretionary wedge have uplifted a series of Neogene flysch basins (van der Lingen and Pettinga, 1980) which define the coastal hills in southern Hawke's Bay.

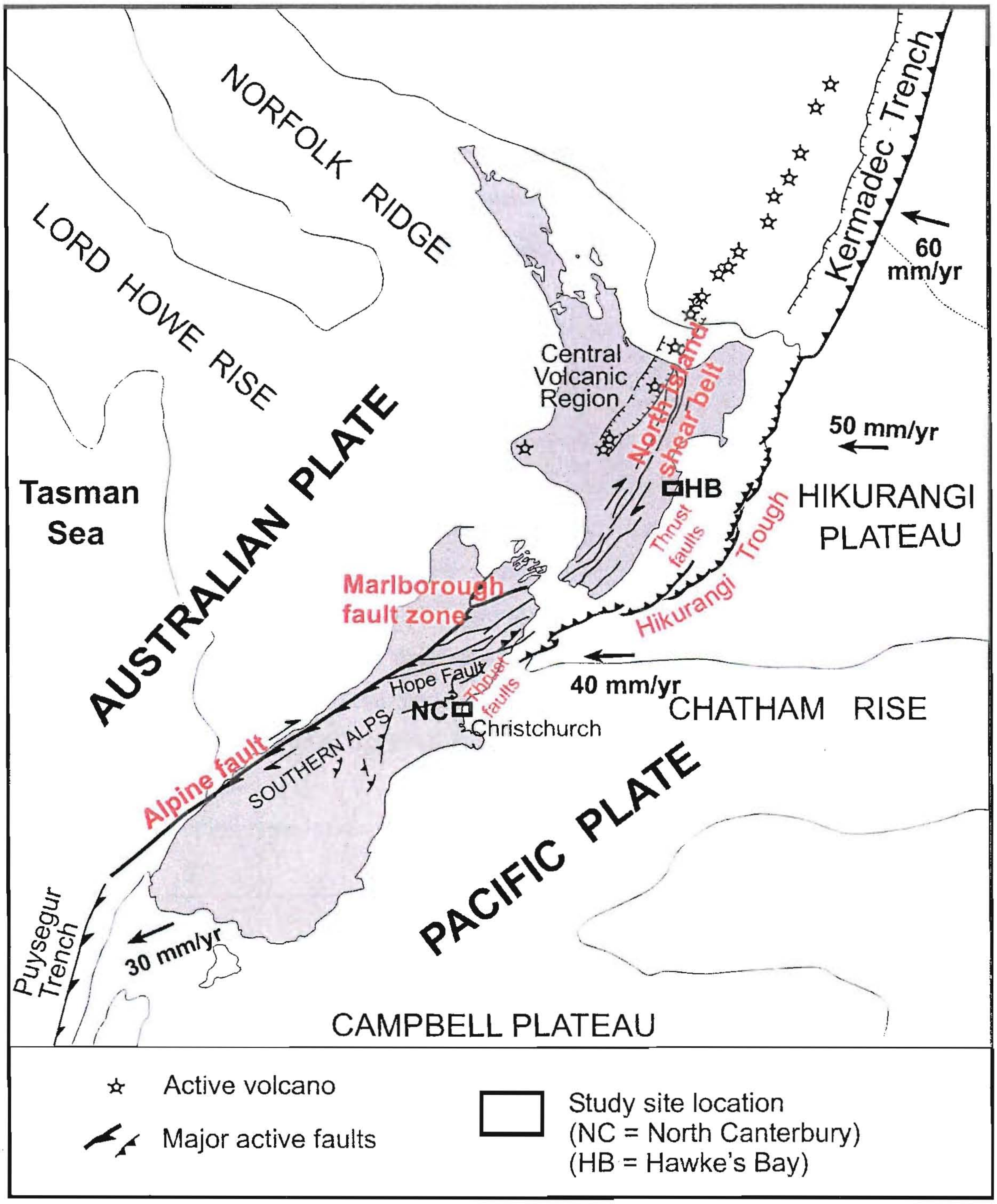


Figure 2.1: Tectonic setting for the New Zealand region, showing New Zealand straddling the Australian – Pacific Plate boundary zone. The major structural and tectonic elements of this plate boundary zone are indicated, and arrows indicate plate motion vectors for the Pacific Plate relative to the Australian Plate. The locations of the study sites for this project are indicated. Figure modified from Pettinga (2001).

Tectonic deformation in northeast South Island is associated with oblique plate convergence and the transition from subduction offshore to tectonic collision and dextral shear onland (Reyners and Cowan, 1993; Barnes, 1996). The Marlborough Fault Zone (Figure 2.1) accommodates a significant amount of this tectonic deformation as the plate boundary deformation is transferred across the northern South Island, from the east dipping Alpine Fault in the southwest to the west dipping subduction zone at the Hikurangi Margin to the northeast. In North Canterbury this transition is associated with tectonic shortening, crustal thickening and uplift and is defined as the North Canterbury Fold and Fault Belt (Pettinga et al., 1998; Pettinga et al., 2001) in which deformation is thought to have commenced some 0.5 – 1.0 million years ago, and which is dominated by both on-shore and off-shore thrust faulting.

Deformation along the east coast of both the lower North Island and upper South Island primarily involves Upper – Cretaceous, Tertiary and Quaternary sedimentary "cover" rocks and indurated sandstone and argillite units of the Mesozoic Torlesse "basement". A significant structural similarity between the two regions is the way in which the sedimentary cover sequences are folded into thrust fault driven asymmetric folds, where thrust fault propagated anticlines develop and synclines form passively in the footwall block, between adjacent anticlinal structures. This results in steep to overturned anticline limbs on the foot wall side of the thrust faults and much shallower dipping back-limbs on the hanging wall side (Figure 2.2). In some situations thrust faults may splay at shallow depths (footwall imbrication), and propagate at shallow dip angles toward the surface.

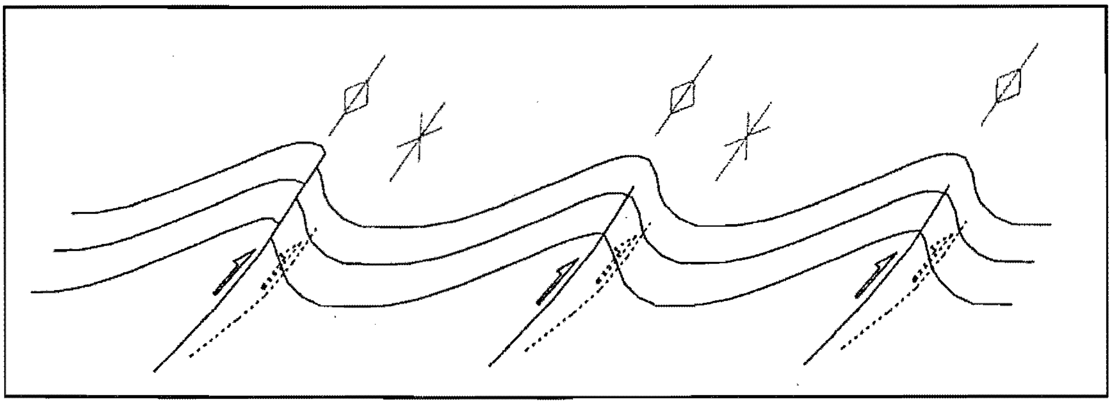


Figure 2.2: Asymmetric thrust fault propagated folding showing anticline (diamond) and syncline (cross) orientations. Schematic representation of folding style in thrust fault zones of east coast of New Zealand (e.g. Lewis and Pettinga, 1993; Nicol et al., 1995; Beanland et al., 1998; Pettinga and Armstrong, 1998; Nicol and Campbell, 2001)

2.3 Site selection

In selecting the two field sites for this project, the following has been considered:

- The presence of deep-seated planar block slides and/or wedge failures. This style of slope instability is widespread in Tertiary soft rock terrain and has a significant influence on catchment morphology and development, yet the issue of why such landslides are able to occur in the natural environment is poorly understood
- The potential for observing the stratigraphically controlled and undisturbed failure plane of any given landslide; and
- The structure and geology should be relatively uncomplicated, with gently dipping sequences of Tertiary soft rock strata, allowing for a high degree of certainty in terms of basin wide stratigraphic correlation.

In order to determine the specific sites for this project, geologic and geomorphologic criteria are considered to be of critical importance. Suitable catchment landscapes for this project are primarily recognisable due to specific geomorphologic characteristics, which are directly controlled by the geology and tectonic setting in which they are developing. The two selected sites are outlined in the following sections.

2.4 North Canterbury site

Kate Valley, North Canterbury is chosen as a study site primarily because of the previously recognised deep-seated Ella landslide which occurs in the middle section of the valley (Geotech Consulting Ltd, 2002). Kate Valley falls within the small, coastal, east draining Kate Stream catchment, which is considered to be representative of the many small catchments developing in Upper Cretaceous and Tertiary soft rock sequences on the east coast of the northern half of the South Island. The catchment is located some 65 km north of Christchurch and about 9 km southeast of Waipara (Figure 2.3).

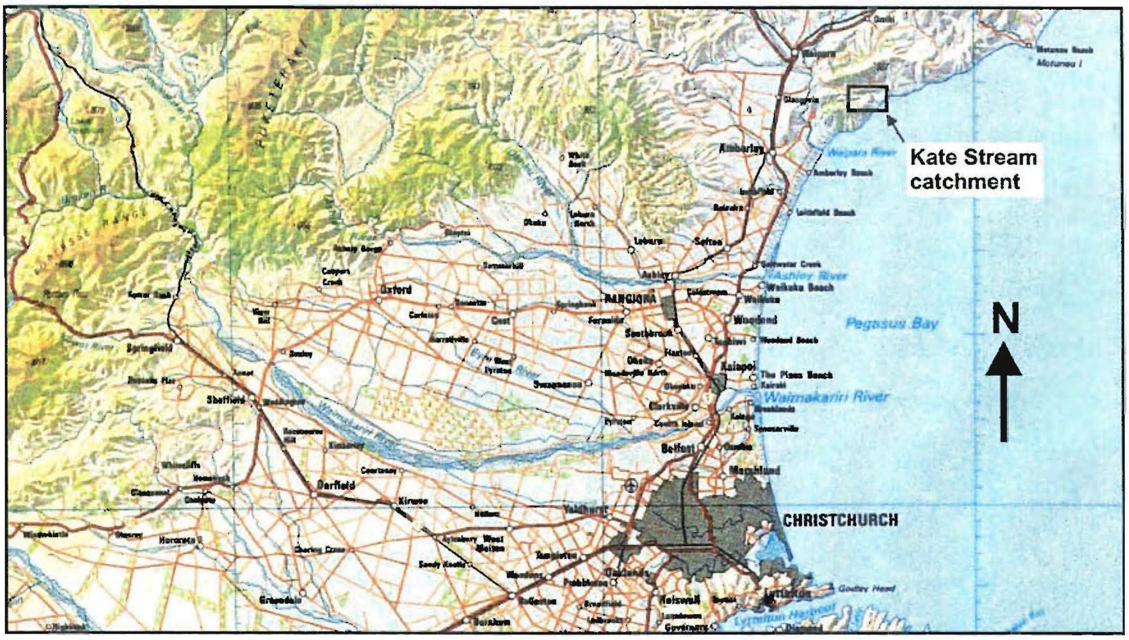


Figure 2.3: Locality map for the Kate Stream catchment shown in relation to Christchurch. Image from TopoMap NZ 1:500k series.

The Ella Landslide is a large (~2 million cubic metres) deep-seated landslide which is inferred to have failed rapidly on an undefined bedding plane surface (Geotech Consulting Ltd, 2002). The landslide has well preserved head scarp and slide block features (Figure 2.4), and the geology and geomorphology of this site fit the criteria as outlined in Section 2.3.

Figure 2.4: View across middle Kate Stream, with the Ella landslide in centre of view showing the well defined head scarp, the main slide block and landslide dam induced alluvial valley fill forming the flats in centre left of view. Photograph taken looking NNE from 5789250N 2497950E (NZMG 260 series topographic map sheet N34).

2.4.1 Geological setting

Structure

Kate Valley lies within the North Canterbury Fold and Fault Belt which is dominated a NE-SW trending structural grain (Figure 2.5). Folding in this zone is driven by southeast dipping thrust/reverse faults deforming the Tertiary cover sequence. Folding is calculated to have accommodated 12-15% of the NW – SE regional shortening from the Hope Fault to the coast (Nicol et al., 1994), and is characterised by strongly asymmetric anticline – syncline pairs in the style indicated in Figure 2.2. The asymmetry in these folds is directly controlled by the upper crustal geometry of fault propagation, and this trend is applicable on a regional scale with major faults traceable for 20 km or more, which in places constrain small terrestrial sedimentary basins (Nicol et al., 1995). Faults which propagate to the surface typically rupture on the steeper northwest limbs, however, many blind thrusts are also contributing to fold development, and faults tend to splay as they near the surface introducing some structural complexity.

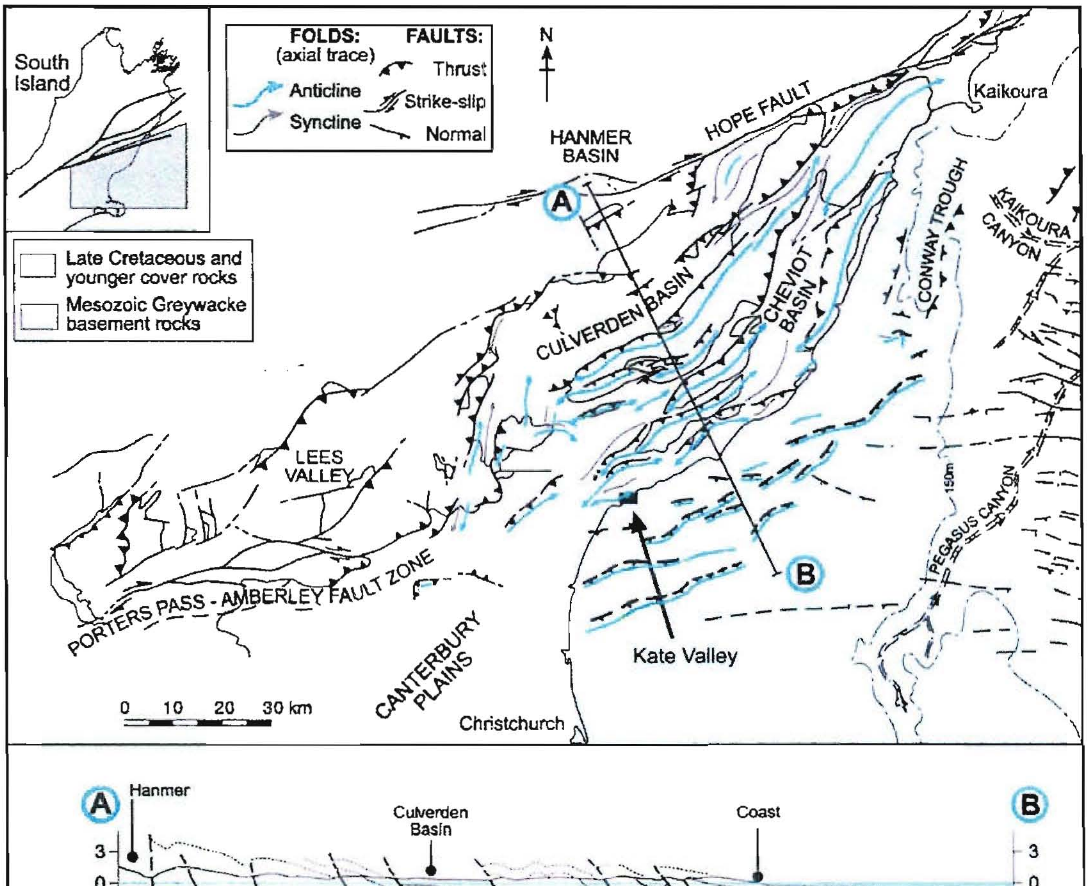


Figure 2.5: Structural setting of North Canterbury showing onshore and offshore fold and fault systems. Cross section A – B shows the relationship of thrust faults to the geometry of the folded cover sequence. The location of the Kate Valley field site is indicated. Modified from Pettinga and Armstrong (1998).

In Kate Valley the structural deformation of the cover sequence is represented by the Teviotdale Syncline – Kate Anticline pair, which are a relatively low amplitude fold pair with dips in and around Kate Valley generally not exceeding 20 degrees (Figure 2.6). While no thrust fault has been mapped in relation to the Teviotdale Syncline it may be inferred that a south east dipping thrust structure of some form occurs at depth.

Stratigraphy

Lithologies surrounding the North Canterbury study area are dominated by Tertiary and Quaternary sedimentary sequences, principally of marine origin, which overlie indurated sandstones and argillites of the Mesozoic Torlesse basement (Wilson, 1963; Field and Browne, 1985). Regional stratigraphy is shown in Figure 2.6, however, only the two formations which are relevant to the Ella Landslide failure will be discussed in the following sections.

Tokama Siltstone Formation

Field and Browne (1985) define the Tokama Siltstone as blue-grey, moderately indurated, calcareous, fine sandy siltstone with scattered shell fragments and calcareous concretions. The material is generally more of a silty fine sand than a siltstone and is finely bedded with occasional (non-silty) sand beds and cemented layers. The formation is assigned an Otaian to Waiauan age (Early Miocene, 13.2 – 21.7 Ma) based on foraminifera. The suggested depositional environment is that of an outer shelf setting (100 – 200 m water depth).

Greenwood Formation

The Greenwood Formation unconformably overlies the Tokama Siltstone but is reasonably similar in character. The Formation may be distinguished by a lack of concretionary layers and a basal cemented shell pebble conglomerate bed up to 5 m thick (Geotech Consulting Ltd, 2002), and the main unit is a blue grey to yellow massive silty fine sand. The formation is assigned a Pliocene to Upper Miocene age (5 – 8 Ma). Field and Brown (1985) infer a depositional setting not dissimilar to today's Pegasus Bay setting with voluminous river systems depositing clastic material interspersed with fine grained deposition in a shallow (< 200 m deep) marine environment.

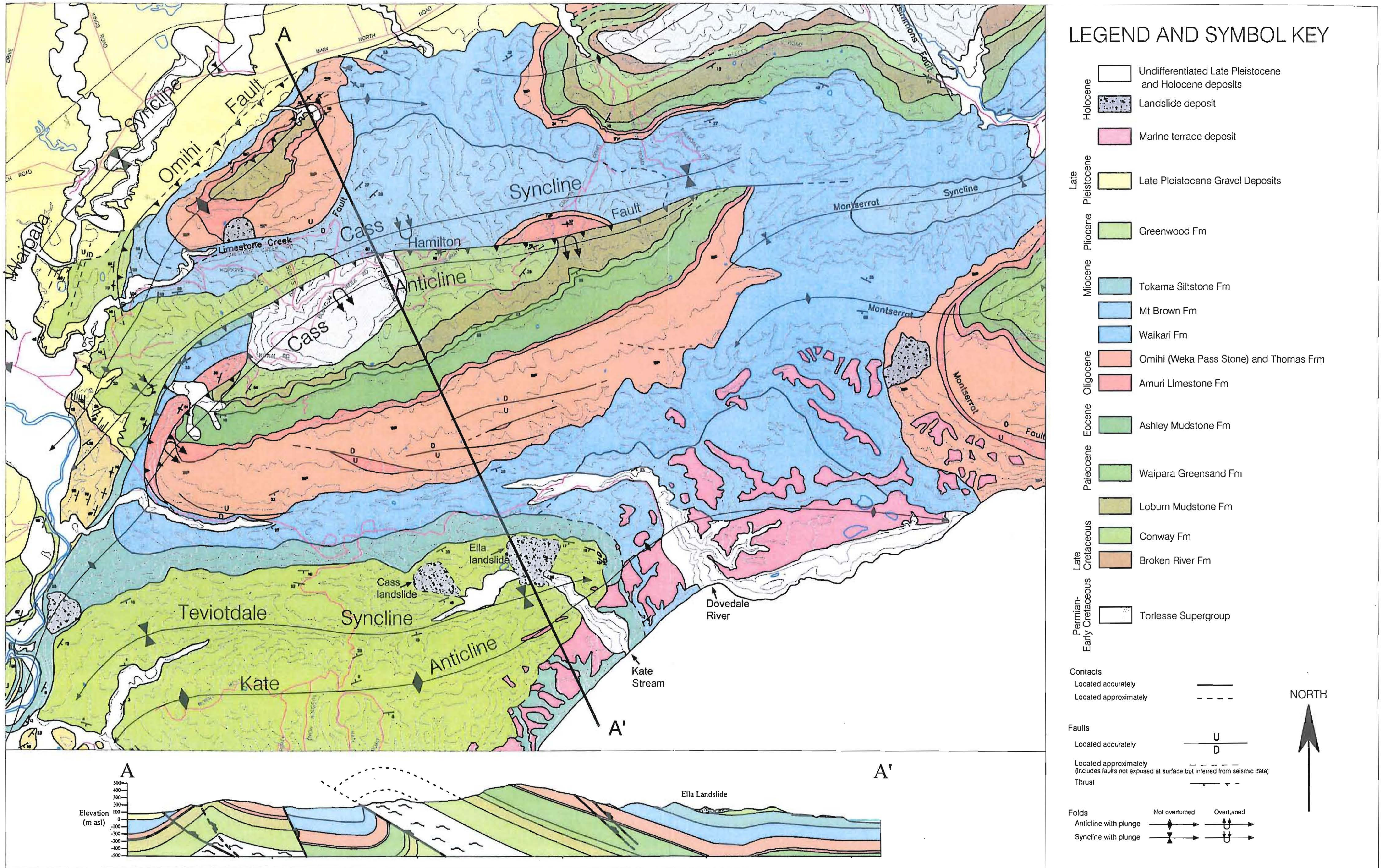


Figure 2.6: Regional geology map for the Kate Stream catchment, North Canterbury. Mapping based on University of Canterbury Department of Geological Sciences, North Canterbury GIS and this project.

Scale 1:50000
0
5 km

2.4.2 Geomorphology of the Kate Valley area

Kate Valley lies within the central and upper portion of a relatively small catchment in coastal North Canterbury. The shoreline along this coast is retreating, which is directly reflected by actively eroding coastal cliffs and advancing wave-cut platform (coastal cliffs are evident at the shoreline in Figure 2.7).

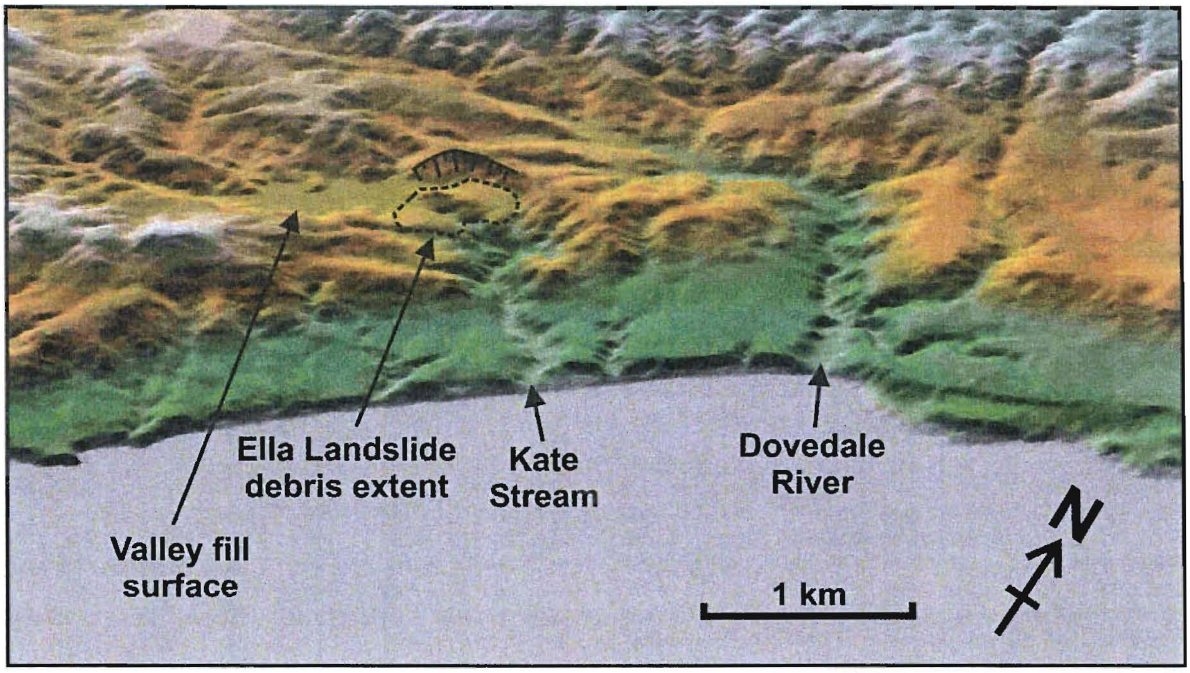


Figure 2.7: Digital elevation model of the Kate Stream catchment showing the location of Ella Landslide. Note the flat area upstream of the outlined landslide debris which defines alluvium accumulated behind the Ella landslide dam. Perspective view is to the northwest at approximately 25° from horizontal.

Ongoing tectonic activity in coastal North Canterbury is reflected by uplifted and sometimes tilted marine platforms (Figure 2.8, and also shown in Figure 2.6 as marine terrace deposits) and deeply incised stream networks. The Kate Stream catchment is representative of the many small catchments developing in the Tertiary soft rock sequences of the North Canterbury fold and fault belt, and the development of these landscapes can be considered in terms of tectonic and climatic forcing (as discussed in Chapter 1).

Influence of tectonic and climatic forcing

Tectonic and climatic forcing have a dominant control on geomorphic development of coastal catchments in North Canterbury. Variation in regional base level, driven by long-term tectonic and climatic forcing in the form of ongoing tectonic uplift and orbitally forced glacial and interglacial cycles, directly influences rates and patterns of stream incision. The marine platforms around the study area (Figure 2.8) have been assigned ages of M1 ≈ 60 kyr, M2 ≈ 80 kyr, M3 ≈ 105 kyr and M4 ≈ 125 kyr (Yousif, 1987) and provide uplift rates of 1.36 – 2.16 mm/yr (Nicol et al., 1994), with a maximum uplift rate for the coastal ranges calculated to be 2.68 ± 0.2 mm/yr. Folding and faulting of Tertiary cover sequences in this area is not thought to have commenced until the early Pleistocene (Nicol et al., 1994), driven by an average absolute shortening rate of 0.8 ± 0.4%/100 kyr in the coastal ranges. In Kate Valley, as in other catchments in the area, asymmetric folding has a controlling influence on topography with the Teviotdale Syncline – Kate Anticline pair being broadly coincident with Kate Valley and the seaward ridge, and Kate Stream in the mid-upper catchment flowing along the axis of the Teviotdale Syncline. Most valley and range systems in the region are aligned NE – SW, reflecting the structural grain of the North Canterbury Fold and Fault Belt.

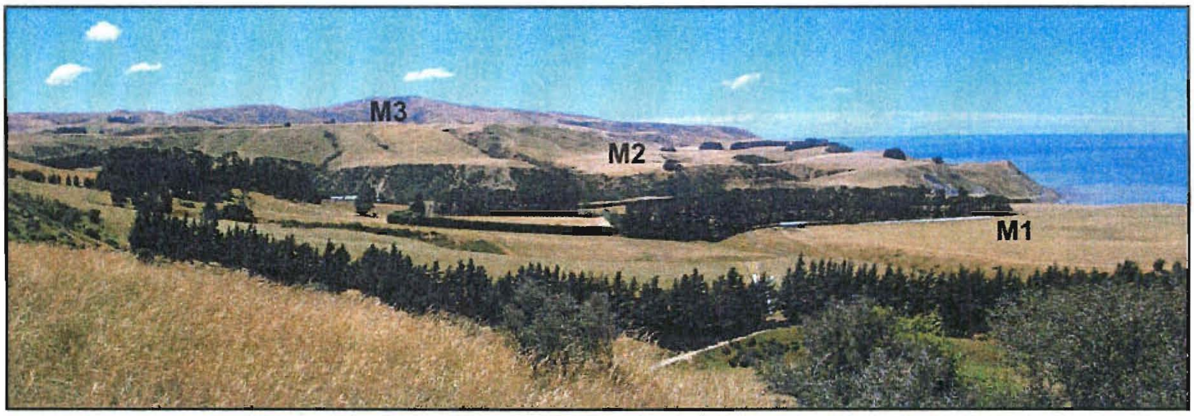


Figure 2.8: Looking north from adjacent to lower Kate Valley towards uplifted marine platforms. The marine platform surface chronology is shown and age control is discussed in text (note surface M4 not visible in this photo).

With the occurrence of rapid base level lowering events stream networks incise deeper into bedrock in catchments and across the exposed shelf and this has a significant effect on the landscape with the initiation of a pulse of accelerated slope instability. In a large catchment where sediment must be transported a significant distance to the ocean, stream response to uplift events may be significantly delayed or smoothed out over a long time as debris aggrades in the river bed. In small coastal catchments the response to base level lowering events is much more rapid as sediment is rapidly flushed out of these systems and streams can immediately begin down cutting into bedrock to attain the new base level. The form of stream incision as a response to a sudden lowering of relative sea-level will typically be by retrogression of reaches of over steepened stream profile gradients or knick points through the landscape (Burbank and Anderson, 2001). Kate Stream and the adjacent Dovedale

River (Figure 2.7) are good examples of streams in coastal catchments which are deeply incised into bedrock and both appear to have established an equilibrium stream gradient for part of their reach. Whereas the Dovedale River stream gradient continues a significant way upstream without sudden variation, the Kate Stream reaches a knickpoint where it encounters the debris of Ella Landslide. Yousif (1987) attributed the barrier to continued incision to active fold growth, however, this interpretation failed to recognize the scale and significance of the Ella Landslide dam on catchment morphology. It is the deeply incised stream valley and subsequent stratigraphic exposure that has enabled this deep-seated landslide to occur on a bedding controlled failure surface. The landslide debris subsequently dammed the stream valley and created a barrier to sediment removal and this effectively disconnects the sediment transport system of the upper and lower parts of the catchment and caused the development of the valley filled alluvial plain evident today (see Figure 2.7).

The deeply incised stream networks mean that the landscape is divided into hillslope areas that are coupled to the present day fluvial system and hillslope areas that have been effectively abandoned, and exhibit paleo-topographic features related to a previous base level regime. Where slopes are coupled to the present-day fluvial system the landscape is characterised by steep-sided valleys and active shallow slope instability, related to the rapidly eroding valley walls which are actively unstable from stream bed to ridge crest. In abandoned areas some mass movement is present, however, the slope form is more rounded and subdued as debris from mass wasting processes is not removed by the fluvial system and forms an apron on lower slopes.

Several instances of deep-seated slope instability are documented in and around the Kate Stream catchment which are inferred to have failed on bedding controlled surfaces (Geotech Consulting Ltd, 2002). The Ella Landslide has been chosen for this study because it is considered to illustrate the influence of a deep-seated landslide on the geomorphic development of this style of catchment, and because critical components for analysing the slope failure (e.g. the failure plane) were considered likely to be exposed as the majority of the post failure topography is well preserved and stream incision has occurred into bedrock and through the lower part of the landslide debris.

Triggering mechanisms for slope failures in this setting can be considered in terms of short-term climatic and tectonic forcing. While the majority of shallow mass movements are generally attributed to high intensity rain storms, it is inferred to be the periodic occurrence of large magnitude earthquake events which initiates large deep-seated landslides. It is the hypothesis of this project that the Ella Landslide was triggered by a large (Mid?) Holocene earthquake with the epicentre in the vicinity of the study area. Further detail on the Ella Landslide morphology, the effect of the landslide on catchment development and landslide triggering will be discussed in Chapters 4 – 6.

2.5 Southern Hawke's Bay site

The Hawke's Bay study area is located approximately 45 km south of Napier (Figure 2.9) and comprises the catchments adjacent to the southeastern margin of the Maraetotara Plateau. This study site is chosen because of the documented widespread occurrence deep-seated slope instability (e.g. Pettinga, 1980, 1992; Pettinga and Bell, 1992) in a landscape predominantly formed on Tertiary soft rock terrain. The Maraetotara Plateau forms an elevated (up to 600 m above sea-level), relatively flat surface along the coastal region of southern Hawke's Bay. The plateau is flanked by numerous deeply incised catchments including the head-waters of the Makara, Ponui and Te Apiti Streams (Figure 2.10), and the geology and geomorphology in these catchments make them ideal study sites in terms of the criteria outlined in Section 2.3.

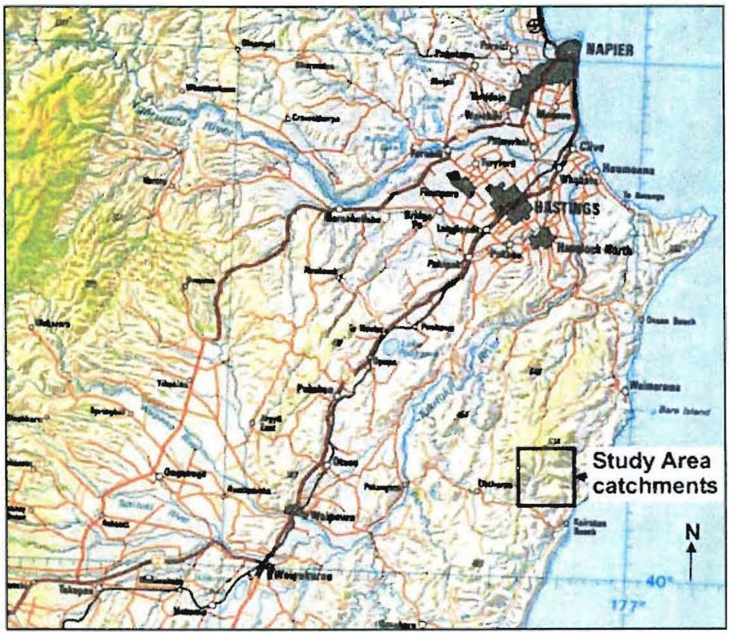


Figure 2.9: Location map for the Southern Hawke's Bay study site, comprising catchments flanking the southeastern margin of the Maraetotara Plateau. Image from TopoMap NZ 1:500k series.

2.5.1 Geology

The Hawke's Bay study site falls entirely within the marine sedimentary sequence of the Neogene Makara slope Basin which has been uplifted and exposed on the now emergent, highest accretionary ridge of the Hikurangi Margin subduction wedge (van der Lingen, 1968; van der Lingen and Pettinga, 1980; Lewis and Pettinga, 1993). The area is characterised by deformed upper Cretaceous to Quaternary strata inferred to be underlain by indurated sandstones and argillites assigned to the Mesozoic basement rocks, in a tectonic province characterised by subduction related thrust faulting and folding.

Figure 2.10: Oblique aerial photograph of the Maraetotara Plateau and adjacent catchments. The main catchments of interest to this study are visible in the foreground with the southeastern Maraetotara Plateau forming the flat area in the middle distance. Photograph taken looking north from approximately 6137000N 2844500E (NZMG 260 series topographic map sheet V22). Photograph courtesy of J. Pettinga.

Structure

The Maraetotara Plateau defines the highest area of the coastal ranges and valleys of southern Hawke's Bay which are included in a structural high formed by an imbricate thrust zone of the emergent part of the Hikurangi Margin accretionary wedge (Pettinga, 1980, 1982; Lewis and Pettinga, 1993). This structural high has been progressively uplifted, and became emergent during the late Pliocene – early Pleistocene (van der Lingen and Pettinga, 1980). Between such structural highs offshore slope basins, including the

Miocene Makara Basin, form by turbidite deposition in landward tilting structural depressions and developed by the bounding northwest dipping thrust faults related to the accretionary wedge development. The Miocene basin fill successions were progressively uplifted along the highest emerging accretionary wedge. The schematic block diagram in Figure 2.11 shows NW dipping thrust faults propagating in the accreting frontal wedge in response to plate convergence and subduction, the setting for basin deposition on the accretionary slope, and the uplifted sedimentary basins and folding within the onshore emergent accretionary ridges. The inboard boundary of the accretionary wedge is defined by the oblique dextral strike slip faults along the east margin of the accretionary trough forming the Heretaunga Basin.

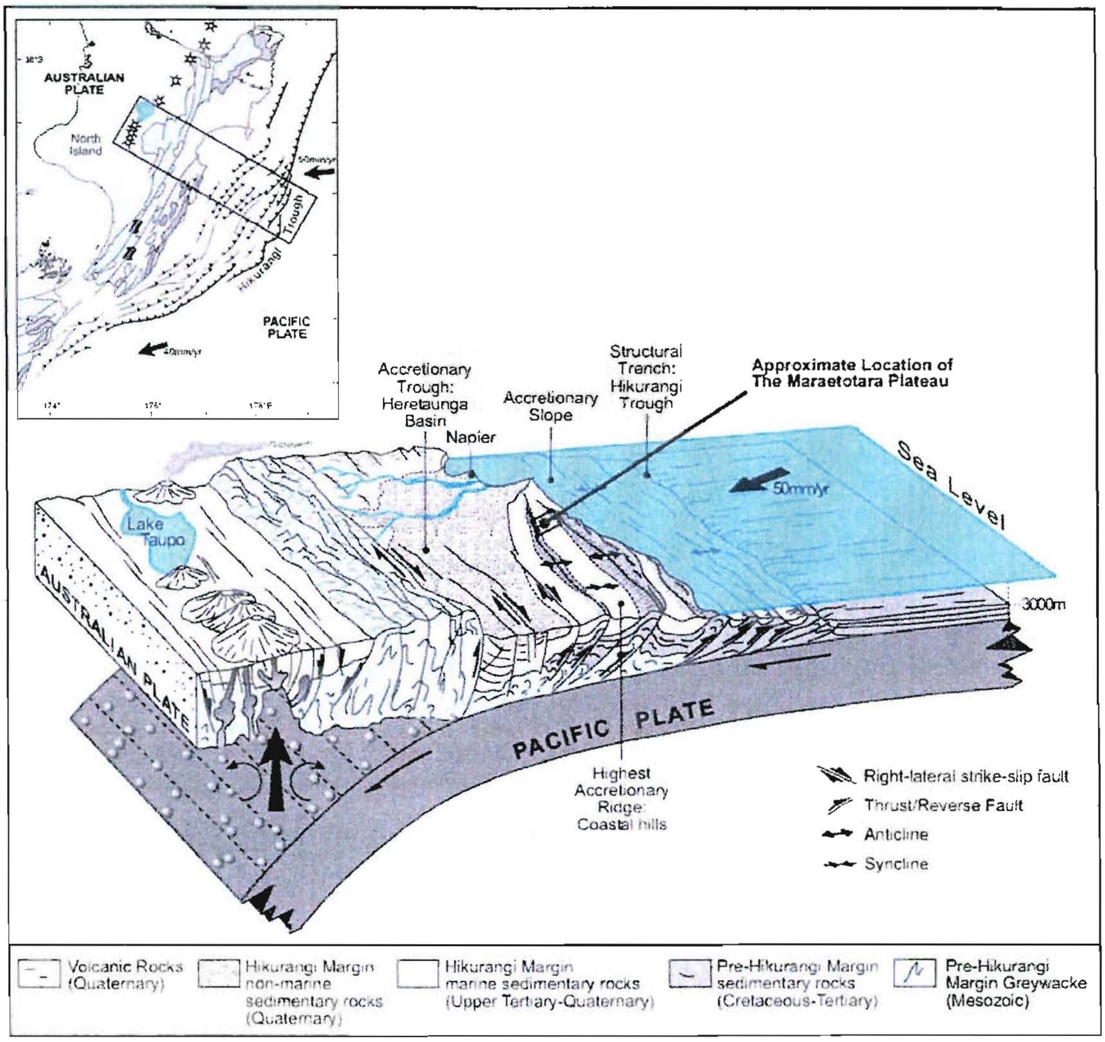


Figure 2.11: Structural setting of Southern Hawke's Bay on the accretionary Hikurangi Margin. The location of the block diagram is indicated with respect to the Australian – Pacific plate boundary (inset). Modified from Pettinga (2001).

As basins become emergent, and possibly inverted, the structural trend of NE – SW striking, northwest dipping thrust faulting and associated mélange zones persist and an overall fold structure of thrust fault driven asymmetric anticline – syncline pairs develops. The exposed Tertiary cover rock sequences on the highest emergent accretionary ridges are characterised by the syncline bound sedimentary basins and intervening narrow structural (anticlinal) highs. The synclines are essentially passive and directly reflect deformation of the adjacent thrust fault propagated anticlines (see also Figure 2.2).

The Maraetotara Plateau is dissected by NE – SW and E – W trending normal faults with relatively minor offsets which have been active since at least the mid Pleistocene (Pettinga, 1980, 1982; Cashman and Kelsey, 1990; Pettinga, 2004). To the east of the Maraetotara Plateau there is a complex zone of east dipping normal faults related to the gravitational collapse of the emergent frontal wedge (Pettinga, 2004). The escarpment on the eastern edge of the Maraetotara Plateau defines the head scarp and western limit of this gravitational collapse, and the erosion resistant limestone which caps the plateau is progressively down faulted to the east and occurs at sea-level near the coast. Faulting within the study area catchments is limited to normal faults related to these extensional zones (Pettinga, 2004). Although frontal wedge related thrust faults have a significant impact on the deformation of the Makara Basin sequence, none of these appear to have been active in the late Pleistocene and Holocene.

Regional stratigraphy

The stratigraphy in the study area is comprised of an Upper Tertiary to early Quaternary marine sedimentary cover rock sequence underlain by an Upper Cretaceous to middle Tertiary sedimentary succession, which is in turn underlain by Mesozoic basement (Figure 2.11). The Makara Basin is one of a series of small (20 by 30 km) uplifted and deformed Neogene sedimentary basins along the east coast of the lower North Island (van der Lingen and Pettinga, 1980). The basin developed as it became bounded by the Waimarama – Mangakuri Coastal High and the Otane Anticlinal Complex during the middle Miocene. As the Hikurangi Margin frontal (accretionary) wedge became emergent it is the Neogene flysch basins which are predominantly exposed at the surface, and hence it is in these that much of the landscape is developing. As the landslides in the study area are only occurring within the Miocene and Pliocene stratigraphy of the marine sedimentary cover rock, only these will be discussed, Figure 2.12 depicts the geologic and structural setting of the study area.

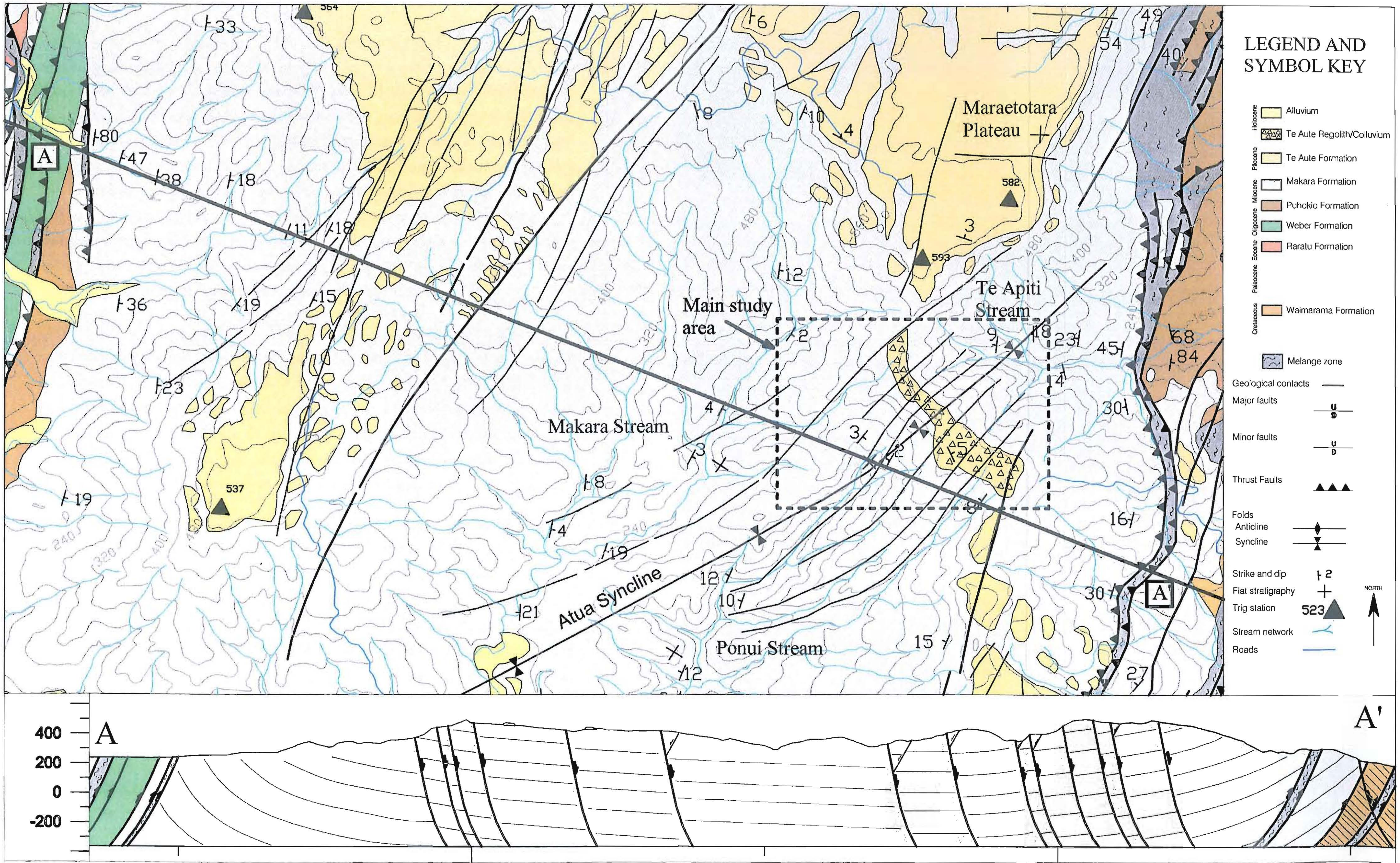


Figure 2.12: Geology of the southeastern margin of the Maraetotara Plateau, southern Hawke's Bay. The inset box indicates the main area of focus within the study area catchments. A geomorphological map of this area is presented in Figure 3.13. Based on mapping from Pettinga (1980) and this study.

Scale 1:25000

0 0.5 1 1.5 2
km

Makara Formation

The Makara Formation is divided into the Motoroa Member of fossiliferous massive sandstones and mudstones and the flysch (alternating sandstone – siltstone/mudstone) dominated Hawea Member (Pettinga, 1980). The landslides included in this study have developed within the Hawea Member of Upper Miocene (Tongaporutuan, 6.5 – 11 Ma) age, comprising alternating finely bedded, graded calcareous sandstone – siltstone and mudstone units. The sediments define typical flysch and subsequently Bouma sequences of sandstone, mudstone and volcaniclastic lithologies (van der Lingen and Pettinga, 1980), and the succession is inferred to have been largely deposited by sediment gravity flows, including turbidity currents, debris flows, slumps and grain flows.

Rhyolitic volcanism was active throughout the development of the Makara basin and as well as discrete tuff (ash fall) beds and sediment gravity flow (reworked) ash beds, the flysch material contains abundant volcanic particles. Offshore, more recent (Holocene) ash layers can be correlated to central North Island volcanic vents (Lewis and Kohn, 1973) and it is inferred that tuffaceous beds in the Makara Formation were deposited in a similar manner throughout the Upper Cenozoic, however, the origin ash fall is more likely to be from the Coromandel volcanic Arc (van der Lingen and Pettinga, 1980).

Despite the descriptor of finely bedded, the main clearly defined (or measurable) bedding occurs at the lower contact of the basal sandstone units of the graded flysch sequences. Bedding surfaces are also found where the thin (typically < 20 mm) un-cemented tuff beds occur, as these define a clear change in lithologic character. Bedding partings occasionally occur within mudstone units towards the upper part of the succession. Overall the Makara Formation is weak, with the siltstone – mudstone units being particularly prone to degradation from wetting and drying cycles (slaking), and the formation as a whole breaks down rapidly with weathering.

Towards the end of the Miocene, increased tectonic activity and uplift terminated flysch deposition and a local unconformity developed. The subsequent shallow marine environment allowed for deposition of the overlying Te Aute limestone in the Pleistocene.

Te Aute Formation

The Maraetotara Member of the Te Aute Formation occurs in the study area and is described as a bedded crystalline coquina limestone and softer calcareous sandstone (Pettinga, 1980). The Te Aute Limestone forms the resistant cap which defines the

Maraetotara Plateau where it occurs with a stratigraphic thickness of approximately 10 m. In places the Te Aute Formation has been eroded away and its previous occurrence is reflected in limestone regolith/colluvium mantling ridge tops and upper slopes.

Wherever it is present the Te Aute Formation hinders catchment progression, as evident by the steep catchment heads associated with its occurrence. The combination of the erosion resistant Te Aute Formation overlying the degradation prone Makara Formation has a significant effect on the geomorphic development of the area.

2.5.2 Geomorphology of the Hawke's Bay site

The Hawke's Bay study site is located on the highest coastal range in Southern Hawke's Bay, and can be broadly described as an uplifted plateau flanked by a dissected landscape of deeply incised catchments (Figure 2.10). The stream catchments under consideration flank the southeastern margin of the Maraetotara Plateau (Ponui, Makara and Te Apiti streams in Figure 2.12) and these are considered to be representative of a greater population of catchments in which lithologically controlled deep-seated landslides are widespread and have a significant influence on landscape development. The geomorphic style of the landscape can again be considered in terms of tectonic and climatic forcing factors (discussed in Section 1.3).

Long-term tectonic forcing (uplift, tilting and folding) has a significant influence on landscape development. Uplift rates in the coastal Hawke's Bay region vary from < 1.0 – 3.0 mm/yr (Lewis, 1971; Pillans, 1986; Beryman, 1993; Litchfield and Berryman, submitted) and these are reflected in the landscape by elevated coastal ranges, uplifted marine terraces and deeply incised stream networks. The deep incision of catchments developing in soft rock terrain is a significant feature of landscapes throughout New Zealand and is controlled by long-term tectonic and climatic forcing causing accelerated base level lowering. Stream incision rates in parts of the east coast North Island have been found to be higher than tectonic uplift rates (Berryman et al., 2000; Litchfield and Berryman, submitted) indicating the importance of tectonic and climatic forcing, as well as other factors such as stream power, rock material and rock mass properties. Controls on stream incision specific to the Hawke's Bay study catchments will be discussed further in Chapter 4.

When a rapid change in relative sea-level occurs, streams in coastal catchments will immediately start adjusting to the new base level. Stream adjustment is commonly passed through the landscape as migrating knickpoints, and previous base levels may be recorded by abandoned terraces (Burbank and Anderson, 2001). At least five base level lowering events can be recognised in a catchment adjacent to this study area (Pettinga, 1980), reflecting ongoing cycles of accelerated base level lowering. In the Maraetotara Plateau and the catchments to the southeast, the landscape can be divided into areas characterised by geomorphic activity. Firstly, where actively incising streams are eroding into catchments and slopes are coupled to the fluvial system from stream to ridge crest (termed rejuvenating landscapes by Pettinga, 1980 and Pettinga and Bell, 1992). Secondly there are areas where erosion was previously active, but is now mostly inactive as slopes are de-coupled from the fluvial system (termed relict landscapes). Relict landscapes generally have a subdued and rounded hillslope appearance and may exhibit relict fluvial erosion features related to earlier geomorphic settings. In rejuvenating areas very steep gully systems reflect active erosional processes, including rock material degradation and shallow slope failure, and accommodate (re)activation of deep-seated slope failures.

Slope failure styles

The spatial occurrence of shallow landslides in the Hawke's Bay study area is considered to be primarily related to regolith production and slope angle, and the triggering mechanism of these is likely to predominantly be high intensity rainstorms. In the rejuvenating landscapes shallow failures typically occur on the steep upper slopes of incised gullies as regolith becomes destabilised at the contact with low permeability (unweathered) bedrock at shallow (< 2 m) depth. The widespread occurrence of shallow slope failure in New Zealand hillcountry, and its consideration as a dominant erosion process (Crozier et al., 1992), is critically affected by historical land-use practice. The clearance of large areas of hillcountry for pastoral farming has had a significant effect on the susceptibility of slopes to shallow regolith/colluvial failure and it is likely that in this study area the occurrence of this style of mass movement has been significantly accelerated since European settlement and deforestation.

Deep-seated landslides are also pervasive in the landscape and must be considered to be a significant, if not dominant factor in geomorphic development in the study area catchments. The spatial distribution and geometry of the deep-seated landslides which occur in the Ponui, Te Apiti and Makara catchments (Figure 2.12) is controlled by critical elements of the Makara Formation rock mass (such as joint/faults sets and critically weak stratigraphic horizons), while the style of deep-seated slope failure varies, and is dependant on bedding dip (which varies between 0 - 25°). The occurrence of periodic large magnitude earthquakes (short-term tectonic forcing) in the area is considered to be the main way in which landslides of such significant size and extent are able to be initiated. This is documented for at least one large landslide in this area (Pettinga, 1987a), and for many deep-seated landslides internationally (e.g. Keefer, 1984). The Amphitheatre landslide is a deep-seated translational block slide at the head of the Ponui Catchment (Pettinga, 1992) which is thought to be representative of a style of slope failure that is widespread in the study area. The controls on deep-seated landslide occurrence, geometry, mode of failure and their influence on catchment development in this study site will be discussed in greater detail in Chapters 4 and 6.

2.6 Chapter summary

Two study sites that have been selected for this project are considered to represent a broader population of landscapes formed in Tertiary soft rock terrains where catchment evolution is significantly influenced by the occurrence of deep-seated bedrock landslides. Both sites are characterised by Tertiary cover rock successions deformed by thrust faulting and thrust fault driven asymmetric folding and dissected by deeply incised stream networks. In both study sites, deep stream incision controlled by tectonic uplift and orbitally forced glacial/interglacial cycles has divided the landscape into areas that can be characterised by geomorphic activity into relict and rejuvenating terrains.

Critical factors within these landscapes which influence catchment development include:

- Rock mass defect and rock material properties which control the geometry, spatial occurrence and mode of failure of deep-seated landslides. Specific properties affecting this are joint/fault set orientations, the occurrence and strength of critical stratigraphic horizons, and rock material durability; and

- Tectonic and climatic forcing factors. Long-term tectonic forcing causes uplift, tilting and folding of stratigraphic successions, while long-term tectonic and climatic forcing influence rapid base level change, causing deep incision of fluvial systems into bedrock.

Chapter Three

3.0 Field investigations and geotechnical testing

The aim of field investigations, including engineering geological and geomorphological mapping and sampling for geotechnical testing, is to collect data which will contribute to both the quantitative and qualitative understanding of rock mass stability, both at catchment and individual landslide scales. The three stages of field investigation include:

Reconnaissance Stage: assess whether or not the proposed landslides/landslide areas will be suitable for analysis (as outlined in Section 2.3).

Main Investigation and Data Acquisition Stage: field investigation in the selected catchments focused on obtaining information required for modelling of discrete slopes, and also for consideration of deep-seated slope stability at a catchment scale. Primary investigation goals include:

- Geomorphological mapping of selected landslides and surrounded slopes
- Mapping of bedrock stratigraphic detail and the rock mass (including the identification of critical stratigraphic horizons which define landslide failure surfaces) to assist with the development of slope stability models and with correlation between deep-seated landslides failing on a common critical stratigraphic horizon; and
- Sampling for laboratory testing of geotechnical properties (including material strength and rock material durability) which influence slope stability.

Model Confirmation Stage: field confirmation of the proposed models for deep-seated slope failure and the concept of *critical stratigraphic horizons*.

The programme of laboratory testing of geotechnical properties was designed to assist with two specific aspects of the project. Of primary importance is an assessment of the strength properties of material that has a controlling influence on landslide stability, and a second area of focus is classification tests which define the origin and nature of these materials.

This chapter is divided into two main sections which discuss investigations undertaken at each of the two selected field sites. The purpose of this chapter is primarily to present data collected during the field and laboratory stage of this study, while geological and geotechnical data synthesis and discussion is included in subsequent chapters.

3.1 Kate Valley, North Canterbury

In the North Canterbury field site (Figure 2.5) the occurrence of the prehistoric Ella Landslide (indicated in Figure 2.6) has had a significant impact on the Holocene geomorphic development of the Kate Stream catchment. The catchment occurs in an area of faulted and folded Upper Tertiary and Quaternary sedimentary strata that have been subject to considerable tectonic uplift in the Late Quaternary which is reflected by uplifted and tilted marine terraces and deeply incised stream networks. Considerable geological mapping has previously been undertaken in this area (Gregg, 1959; Wilson, 1963; Yousif, 1987; Geotech Consulting Ltd, 2002), and the regional geology is presented in Figure 2.6. The Kate Stream catchment is selected as representative of the numerous small coastal catchments in this area that are developing in response to tectonic and climatic forcing, and specifically where catchment development is being significantly impacted by deep-seated slope failure. Field investigations within the Kate Stream catchment were primarily carried out in the early part of 2004.

3.1.1 Field Mapping

A preliminary visit to the North Canterbury field site indicated that with further field investigation it might be possible to observe the landslide failure plane in an in-situ condition within stratigraphy exposed by stream incision directly downstream of Ella Landslide debris. Observations (based on slide block geometry) during this site visit concurred with the hypothesis that the landslide could have failed as an intact block slide on a single, discrete bedding plane controlled failure surface. Subsequent to the reconnaissance visit, detailed engineering geological and geomorphological mapping was undertaken to define the geomorphic processes occurring within the catchment, the morphology, extent and slide block geometry of the Ella Landslide, and finally to document the stratigraphy pertinent to the occurrence of this slope failure. Mapping was undertaken using vertical aerial photograph interpretation (NZ Aerial Mapping Ltd run 1824/48-55 flown in 1953) and field mapping which enabled the development of a geomorphological map of the mid-lower Kate Stream catchment (Figure 3.1).

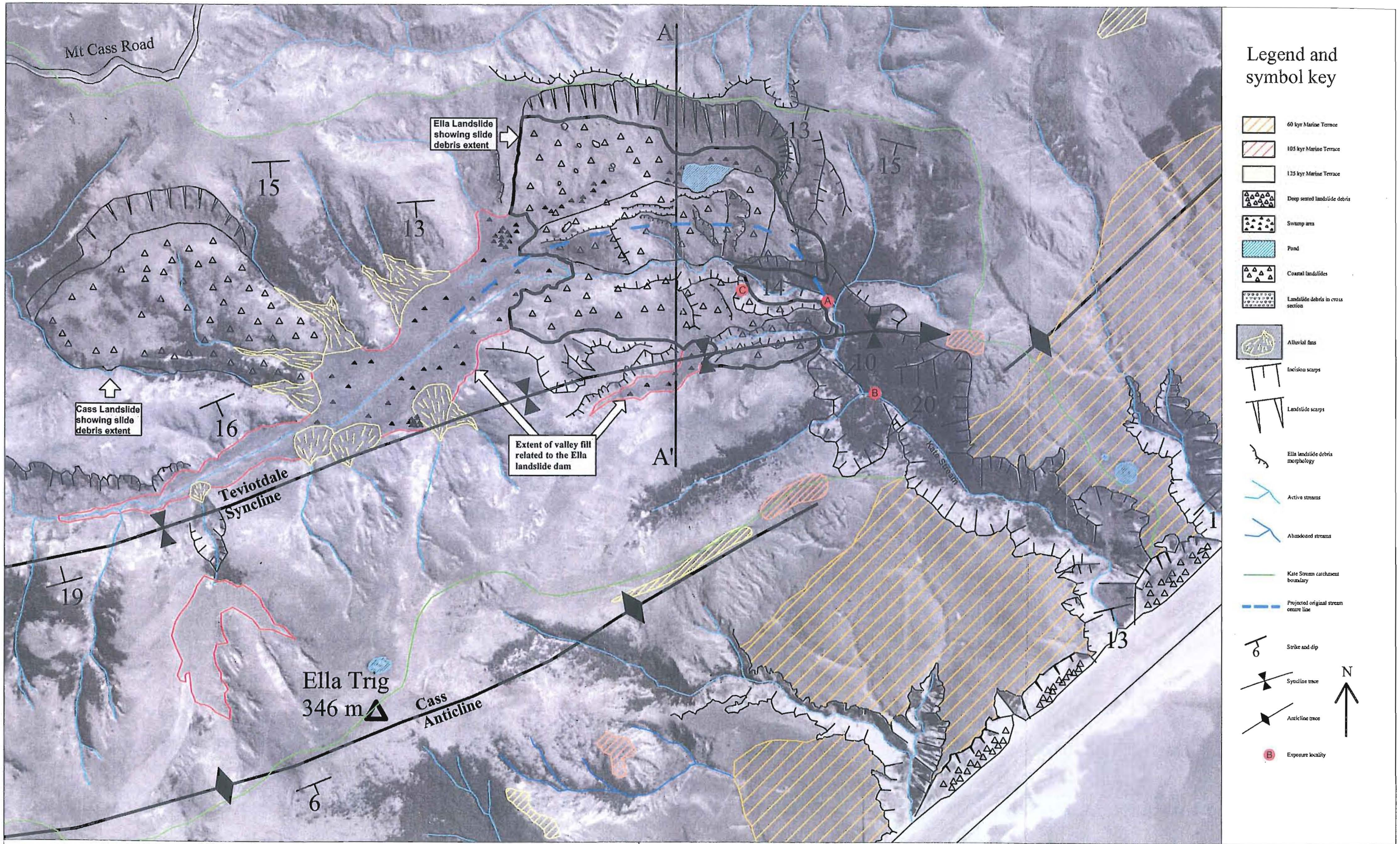
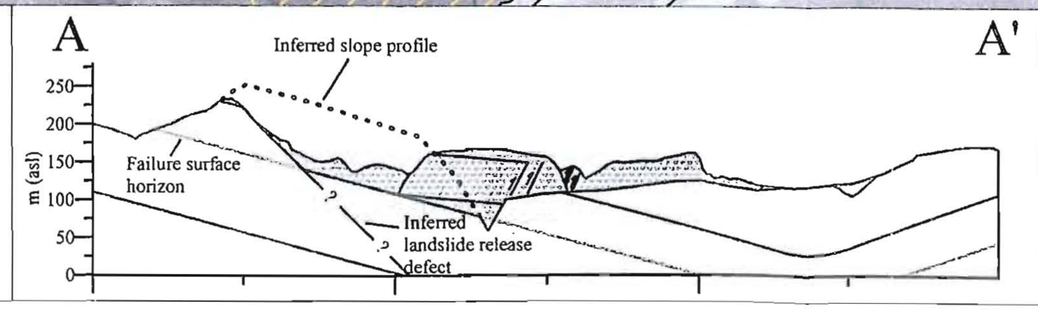


Figure 3.1:
Geomorphological map of Kate Stream catchment. Cross section A-A' shows the failure geometry for Ella Landslide and slide block debris and the inferred cross valley profile. Exposure localities "A", "B" and "C" are discussed in text.



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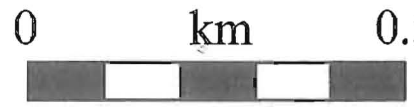

Field mapping revealed a thin, pre-sheared, clay rich layer, which may be considered to define a critical stratigraphic horizon within the Tokama Siltstone Formation, occurring within in-situ stratigraphy but in close proximity to landslide debris (exposure locality "A" in Figure 3.1). Mapping and the identification of this horizon confirmed that Ella Landslide would meet project requirements (outlined in Section 2.3), namely that the landslide has failed on a bedding plane surface which can be observed in-situ in the stratigraphy. Furthermore, the deep-seated slope failure has significantly affected catchment development. Subsequent to the discovery of the critical stratigraphic horizon, the lateral stratigraphic extent of the horizon was investigated. While it was not possible to confirm the lateral continuity of the horizon more than several metres past the initial exposure, a similar thin, clay rich, pre-sheared horizon was observed exposed in the stream bed less than half a kilometre downstream (exposure locality "B" in Figure 3.1). While it is considered to be possible that this second exposure of a thin, clay rich, pre-sheared horizon would correlate stratigraphically to the initially observed exposure, this has not been confirmed.

3.1.2 Landslide failure surface investigation

To assess the origin of the critical stratigraphic horizon observed in the Tokama Siltstone, which is inferred to form the failure surface of the Ella Landslide, physical and mechanical characterization tests have been carried out to address both the stratigraphic position and origin (lithological vs mechanical) of the clay rich horizon. These tests include: i) grainsize analysis of material in and around the horizon; ii) clay mineralogy of the material in and around the horizon using X-Ray diffraction; and, iii) micro fabric assessment of material in and around the horizon using a scanning electron microscope.

Grainsize analysis

Grainsize analysis has been undertaken to enable comparison of the grainsize of clay material within the critical stratigraphic horizon with material occurring stratigraphically adjacent to the horizon. Figure 3.2 shows the exposure where sampling for grainsize analysis was undertaken and a close up of the critical stratigraphic horizon (inset left). Continuous sampling was undertaken for approximately 0.5 m stratigraphically above and below the critical stratigraphic horizon in 20 mm increments, using a putty knife to cut a vertical channel and remove samples in bedding parallel slices.

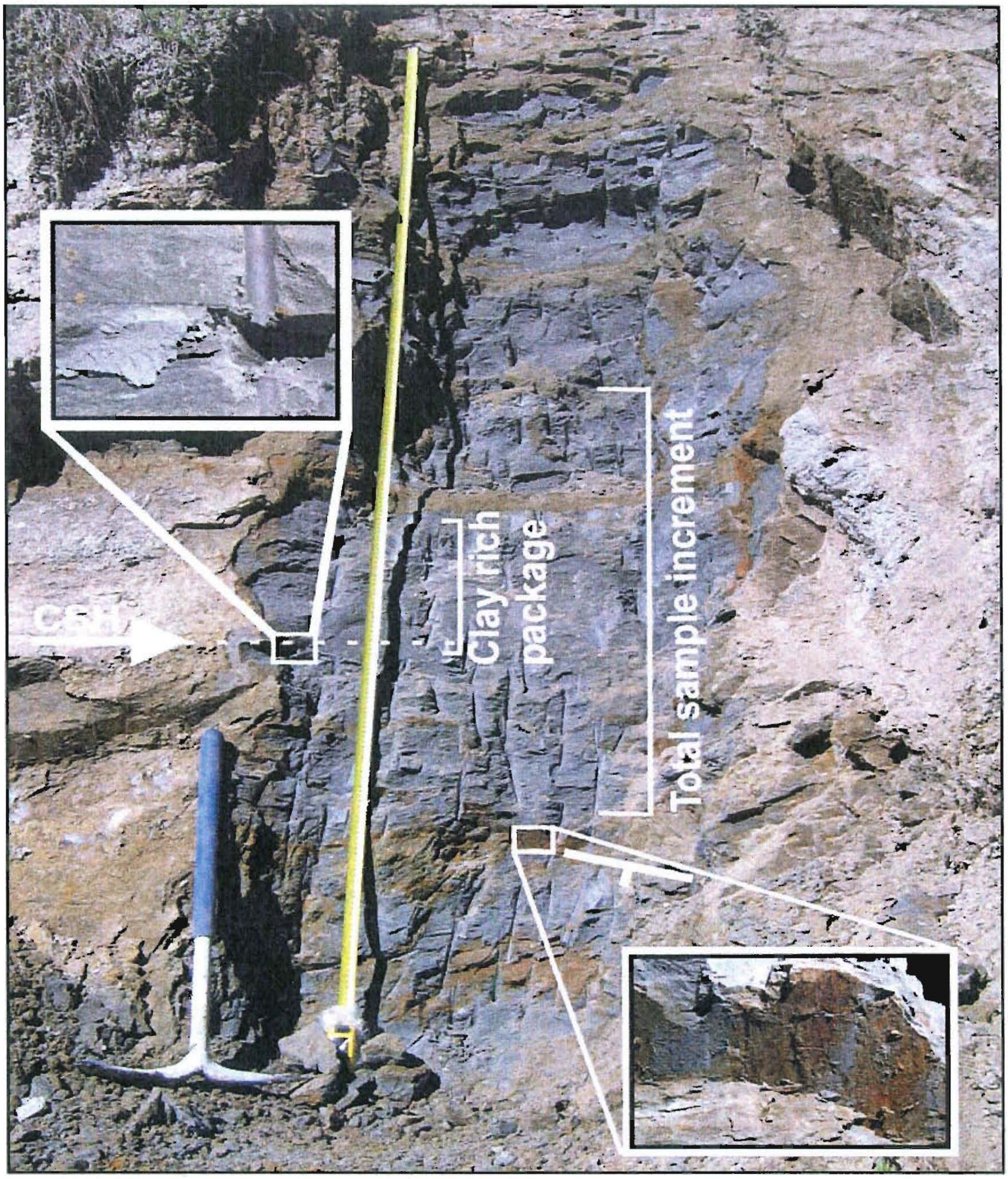

Figure 3.2: Photograph of the stratigraphic exposure of the critical stratigraphic horizon inferred to be the failure surface of Ella Landslide. The location of the critical stratigraphic horizon (CSH) is indicated by the arrow at the left and a close up is shown in the inset above as the (~5mm thick) dark horizontal layer. A defect is indicated and shown in close up (inset). The location of the exposure is shown as exposure locality "A" in Figure 3.1. The geopick is 0.63 m long.

Minor weathering of the Tokama Siltstone has occurred at this location, and is reflected by iron oxide staining on exfoliation defects and in some sand rich layers. This is not considered to affect sample quality, however, as none of the oxidised sand layers occur within the sample interval and approximately 150 mm of surface material was removed prior to sampling to avoid exfoliation defects and ensure that material was not desiccated from surficial drying in outcrop exposure.

For grainsize analysis the Micrometrics® Saturn DigiSizer 5200 (hence referred to as a DigiSizer) is used. This method measures light scattered from particles based on their size, shape, refractive index and the wave length of incident light (Micrometrics Instrument Corporation, 2003). The method allows a sample with a grainsize range of 0.1 to 1000 micrometers to be analysed in approximately 30 minutes, repeating the test 3 times (or as decided). No published reference material has been found which discusses the validity of the DigiSizer, however, work at the Desert Research Institute in Nevada, USA (T. Caldwell, pers. comm. 2004) has found that the DigiSizer reproduces traditional pipette analysis statistically well (within R^2 of 0.9) for their samples. The samples analysed by the Desert Research Institute contain more than half medium to fine sand and the results can underestimate silts and clays by 5% or more, however, due to the predominantly fine grained nature of the samples analysed here this method is considered to be adequate.

The DigiSizer requires that you set the beam obscuration (related to sample grainsize and its concentration in liquid) based on the expected mean particle diameter. This creates an increased error when the sample is well graded as the obscuration can only be focused on one grainsize range within the sample (bins are < 1 μm, 1 to 10 μm, 10 to 100 μm, 100 to 1000 μm). As the grainsize of interest to this study is the relative concentration of fine grained material the obscuration was set to minimise the error for this fraction, and three tests were conducted from each sample, giving nine analyses per sample (Appendix I shows all analysis curves). Variation between test runs is generally in the order of 0.1 % volume frequency, except for some well graded samples where the coarser fraction may show more variability. As the grainsize of interest to this project is the clay size fraction, the obscuration values are targeted at this particle size increment so that the results for the clay fraction can be considered as accurate as possible. As the purpose of grainsize analysis is to compare samples and exact grainsize is not considered crucial, the level of accuracy achieved is deemed sufficient.

The selection of the grainsize curves in Figure 3.3 shows that the clay horizon is clearly finer grained than the surrounding material. The upper and lower samples were taken just outside the contact of the clay rich package in which it occurs (indicated in Figure 3.2) and shows that grainsize increases away from the clay horizon.

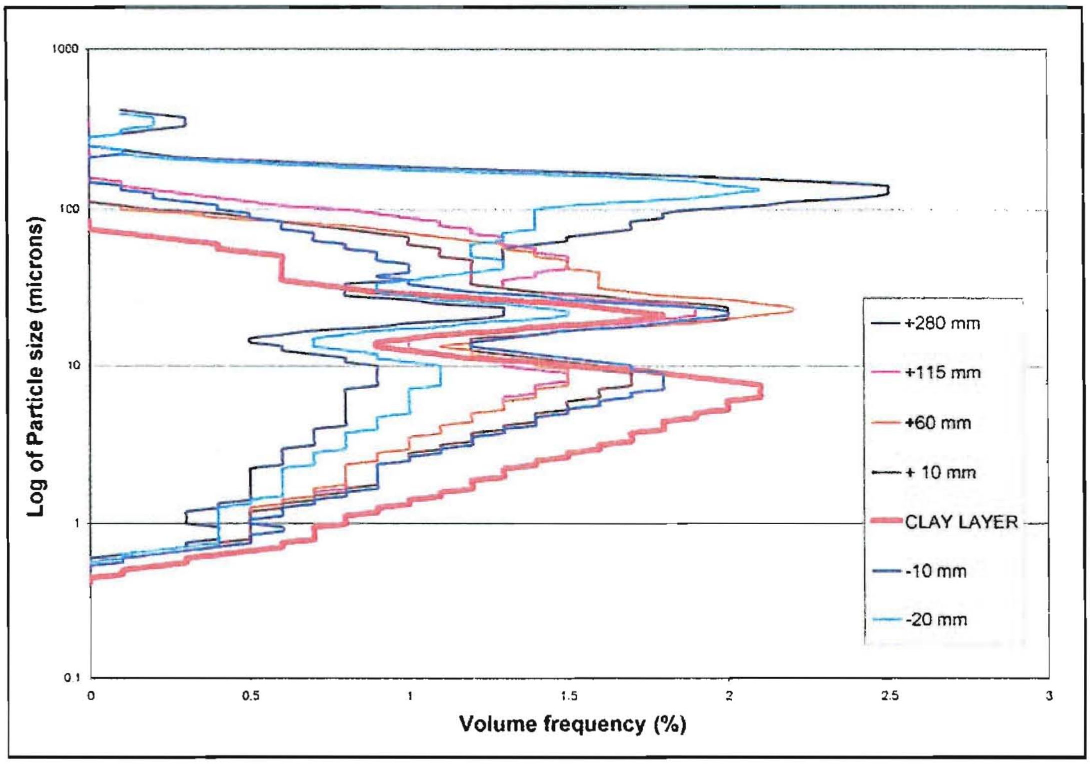

Figure 3.3: Selected grainsize curves for comparing the Tokama Siltstone critical stratigraphic horizon (clay layer) grain size with stratigraphically adjacent material. Samples obtained from exposure shown in Figure 3.2.

Clay mineralogy – X-Ray Diffraction analysis

X-Ray Diffraction (XRD) is used to determine which, if any, clay minerals are present in a given sample (for procedural details see Moore and Reynolds, 1989). In this study XRD is used to compare the clay mineralogy of material within the critical stratigraphic horizon to that stratigraphically either side of it (Figure 3.4). Samples taken from within and directly above and below the critical stratigraphic horizon (±10 mm in Figure 3.3) show that the critical stratigraphic horizon contains a significantly higher concentration of the clay mineral Kaolinite than adjacent material.

Clay mineralogy – Scanning Electron Microscope analysis

The microstructure of the Tokama Siltstone (at individual grain scale) might provide information regarding controls on the location of the critical stratigraphic horizon. Using the Scanning Electron Microscope (for procedural details see Reed, 1996) individual scanned images were taken of material within the critical stratigraphic horizon, and of material either side of it (Figure 3.5). An intact sample was also scanned to examine an oriented surface at the contact of the intact Tokama Siltstone with the critical stratigraphic horizon (Figure 3.6) for an indication of shear direction on this surface.

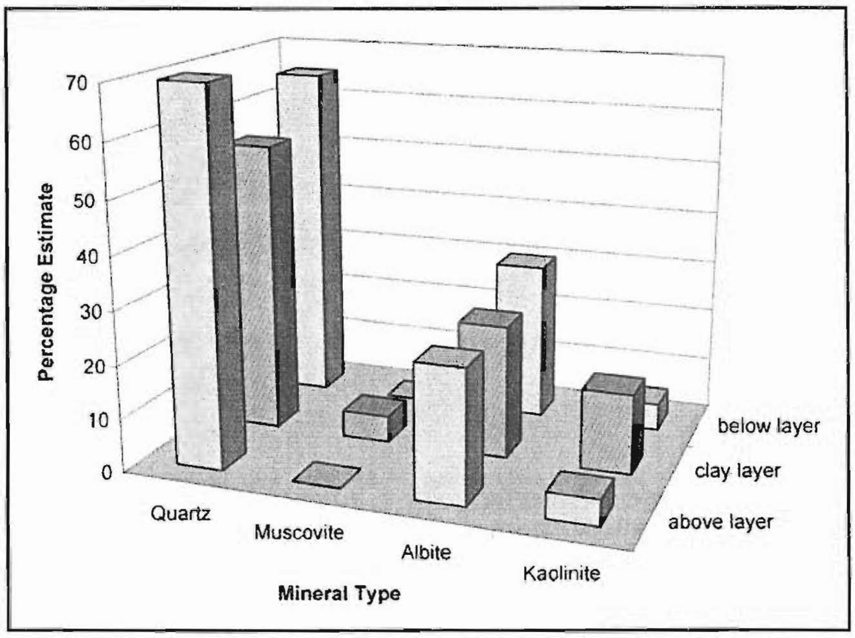

Figure 3.4: X-ray diffraction analysis results for the Tokama Siltstone critical stratigraphic horizon and material stratigraphically either side. Samples obtained from exposure shown in Figure 3.2.

Image B in Figure 3.5 shows that clay minerals are aligned which confirms that there has been shear within this layer, by comparison particles images A and C appear to be randomly oriented. Figure 3.6 shows that there is some orientation to the shearing (diagonally bottom left to top right) but a sense of displacement direction is not conclusively indicated. For further discussion of the significance of these images refer to the following chapter.

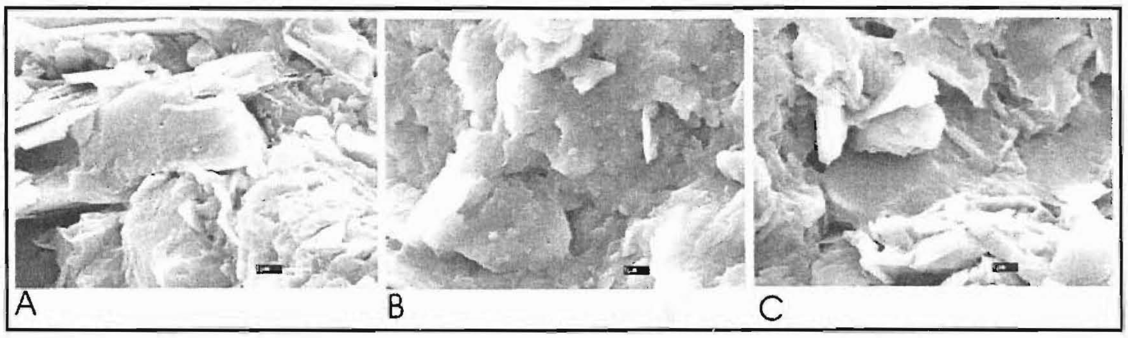

Figure 3.5: Comparison of microstructure immediately below (image A), within (image B) and immediately above (image C) the Tokama Siltstone critical stratigraphic horizon. Samples obtained from exposure shown in Figure 3.2

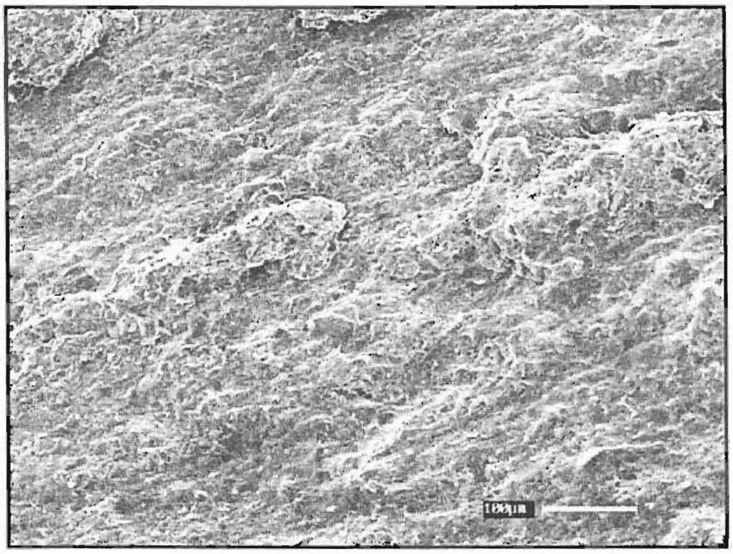

Figure 3.6: Scanning electron microscope image of micro-slickensides occurring on the top surface of the Tokama Siltstone critical stratigraphic horizon. The scale bar is 100 microns. Sample obtained from exposure shown in Figure 3.2

3.1.3 Material strength assessment

A main objective of geotechnical testing is to quantify the pertinent strength parameters of the materials involved in the Ella Landslide slope failure. Rather than embarking on a fully comprehensive programme of strength testing it is considered that the most time and resource efficient approach is to consider those properties that had a controlling influence on the slope failure and focus on quantifying them in detail. For the relevant geological formations to this study there are considerable geotechnical data available from the investigations carried out by Geotech Consulting Ltd. (2002), and these are presented in Table 3.1.

Residual vs. peak strength

Peak strength is the maximum sustainable shear stress for a given material (Barnes, 2000), and is the strength realised when the original material fabric is tested. The residual strength is that mobilised after significant strain causes particle rearrangement and is the lowest shear strength that will be realised by a material.

The Ella Landslide is an intact block slide which has a failure surface defined by a thin and laterally continuous horizon in the bedded Tokama Siltstone. Pre-shearing of this critical stratigraphic horizon (see Figure 3.5) indicates that it is at or very near to residual strength, and hence the defining strength for the stability of the pre-failure slope is considered to be the residual strength value for this horizon.

	Tokama Siltstone				Greenwood Formation			
Grainsize	**Clay (%)**	**Silt (%)**	**Sand (%)**		**Clay (%)**	**Silt (%)**	**Sand (%)**	
	7-10	32-40	50-61		5-8	30-38	54-65	
Character	**LL**	**PL**	**PI**		**LL**	**PL**	**PI**	
	26-44	20-29	13-21		25-34	18-20	14-15	
Strength	**θ**	**C**	**θ'**	**C'**	**θ**	**C**	**θ'**	**C'**
Weathered	27°	98kPa	38°	118kPa	39°	44kPa	42°	106kPa
Un-weathered	38	233	37	176	Not Available			
Permeability	$\leq 1.28 \times 10^{-8}$ ms^{-1}				1.61×10^{-8} ms^{-1}			

Table 3.1 Collation of geotechnical testing data available for the Tokama Siltstone Formation and Greenwood Formation sampled in upper Kate Valley, from Geotech Consulting Ltd (2002).

Justification for chosen test

Fell et al. (1987) consider that where possible strength testing of pre-existing shear planes should be carried out directly on samples of relevant planes of weakness retrieved from intact stratigraphy by direct shear testing of intact block samples. If this is impractical then the ring shear test for remoulded material should be used to define the strength of gouge material. Where the existing plane of weakness contains a discernable thickness of gouge material at residual strength, it could be argued that all shear resistance will occur in that gouge material. In this situation there should be no disadvantage in testing the gouge directly, rather than focusing on the contact of the gouge and intact material. In the Tokama Siltstone critical stratigraphic horizon there is some indication that the shear displacement has concentrated along one contact of the gouge material, as indicated by a flat upper contact, while the lower contact has an undulating and irregular surface. Despite this, it is considered that the minimum shear resistance would have occurred within the gouge material when the slope failed, and hence the defining strength of the critical stratigraphic horizon is the residual strength of gouge material, rather than the shear strength of the contact with intact Tokama Siltstone.

The Tokama Siltstone is amenable to having an intact block cut and transported to the laboratory for geotechnical testing as the siltstone is of low strength and is able to be excavated by hand. Due to known problems testing intact soft rock in both soil and rock shear testing devices (Mackey, 2003), and following consideration and discussion with respect to the benefits of intact block sample testing versus remoulded gouge testing (J. Pettinga and T. Davies pers. comm. 2004), it was subsequently decided that ring shear testing of the remoulded gouge material within the critical stratigraphic horizon would provide sufficient results for the study.

Sampling methodology

Sampling of the critical stratigraphic horizon within the Tokama Siltstone for strength testing was undertaken from the same exposure and at the same time as detailed sampling for grainsize analysis. An approximately 300 x 100 mm area of the critical stratigraphic horizon was uncovered and the plastic clay layer removed and sealed in bags to retain field moisture content.

Ring shear testing method

The details of the ring shear test device and its operation are described by Harris and Watson (1997) and full procedural detail will not be discussed here. As samples for ring shear testing are suggested to be fully remoulded, the field moisture content is not retained during the test and the moisture content is recommended to be equal to or less than the plastic limit.

Initial operation of the ring shear device failed to produce an obvious shear plane through the sample, and as shear could be occurring at the sample-platen interface, it was considered that the test results would not be valid. Modification to the upper platen of the ring shear was undertaken by affixing a layer of 80 grit wet and dry sandpaper (Figure 3.7).

The rational behind this modification was that with the increased roughness on the platen surface no slip would be able to occur on the interface and rupture would be forced to occur within the clay material. The residual strength of a soil is not unique for a given sample (Hawkins and Privett, 1985), but rather depends on the normal stress under which the sample is tested. Figure 3.8 shows a generic curve for what they refer to as the "complete failure envelope", where the flat section of the curve typically occurs at normal stresses above 200 kPa.

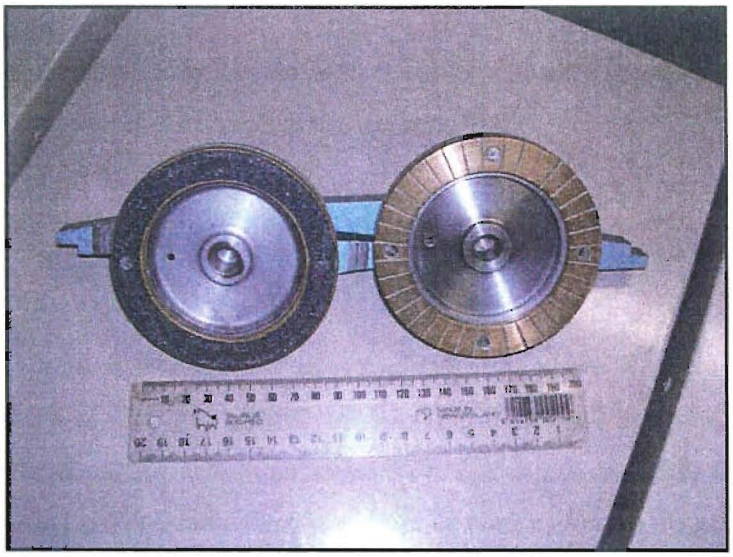

Figure 3.7: Modified ring shear platen (shown on left). The platen on the left has had 80 grit sandpaper attached to the face.

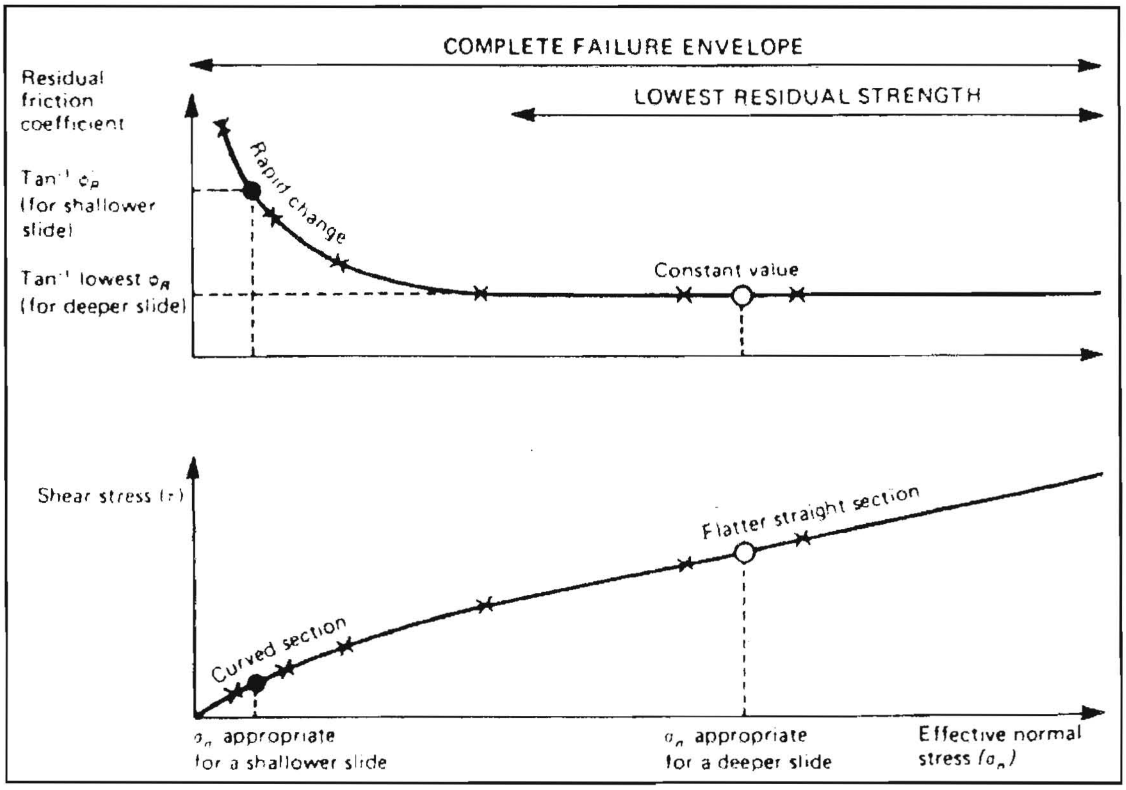

Figure 3.8: The complete failure envelope of Hawkins and Privett (1985).

The appropriate residual strength value for a given situation will depend on the overburden pressures acting on the material in the slope in question. The gradient for the failure envelope curve for residual strength of a soil flattens above 200 kPa, and it is values from this part of the curve that are applicable to slope stability analysis of deep failure.

Ring shear test results

Two ring shear tests have been carried out on the critical stratigraphic horizon material from the Tokama Siltstone, with each test consisting of at least three stages of increasing normal stress (Figure 3.9).

The linear relationship between the points above an effective normal stress of 200 kPa gives confidence that the strength results are valid. The lower data set shows the first point falling below the trend line used to define the effective angle of internal friction (Θ'_R, defined by the slope of the line) and the cohesion (C'_R, defined by the Y-axis intercept). This point represents testing at normal stress below 200 kPa and it is thought that this represents the curved part of the complete failure envelope (Figure 3.8), and as the trend line passes exactly through the centre of the other three points it is considered that these are valid test results.

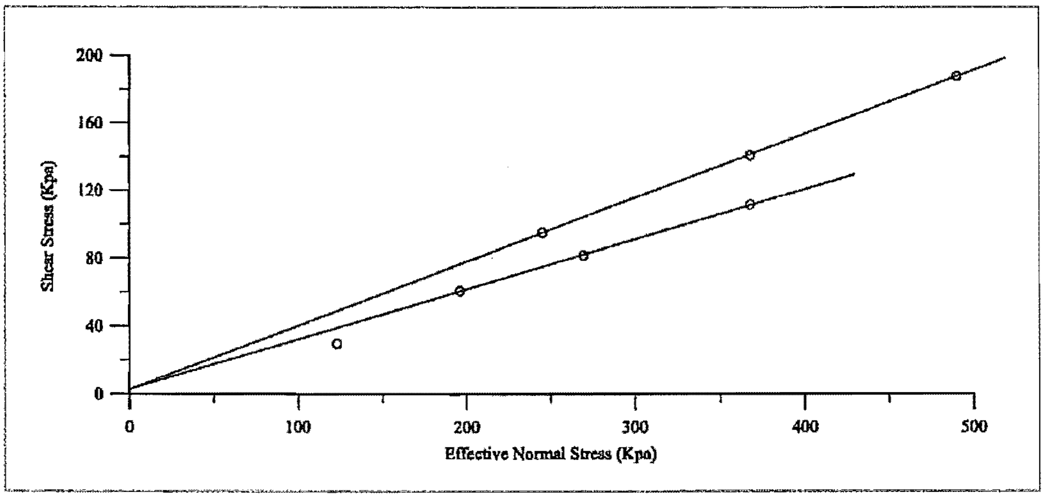

Figure 3.9: Data from the two ring shear tests carried out on the Tokama Siltstone critical stratigraphic horizon

There is a 5° difference between the Θ'_R values obtained from the two tests. The reason for this discrepancy is not clear, the moisture content for the two samples was within 2%, and the test procedure was identical. There may have been some variation between samples that could cause this to occur, such as variation in coarser particle content or the concentration of clay minerals. Both tests indicate a low cohesion value of $C'_R = 2.6 - 2.7$ kPa, and $\Theta'_R = 16 - 21°$. While these values are not particularly low for residual strength values of clay rich material, when compared to the intact strength of the Tokama Siltstone

(C'_R = 176 kPa, and Θ'_R = 37°) this is clearly a very weak horizon within the stratigraphy, and therefore critical to slope stability.

3.1.4 Joint set investigation

Geotech Consulting Ltd. (2000) considered the Tokama Siltstone Formation to be massive and undeformed with essentially no defects in the form of joints and during field investigations encountered only one defect dipping at 45° with slickensides raking 60° to the horizontal. Field investigations during this study also failed to find significant joint sets within the vicinity of Ella Landslide, however, the few relatively persistent joints observed and measured are presented in the stereographic plot in Figure 3.10, and a slickensided defect observed near the critical stratigraphic horizon is shown in Figure 3.2. Given the tectonic setting in which this study site occurs it might be expected that more rock mass defects would be present within the Tokama Siltstone Formation and the significance of this will be considered in Chapter 4.

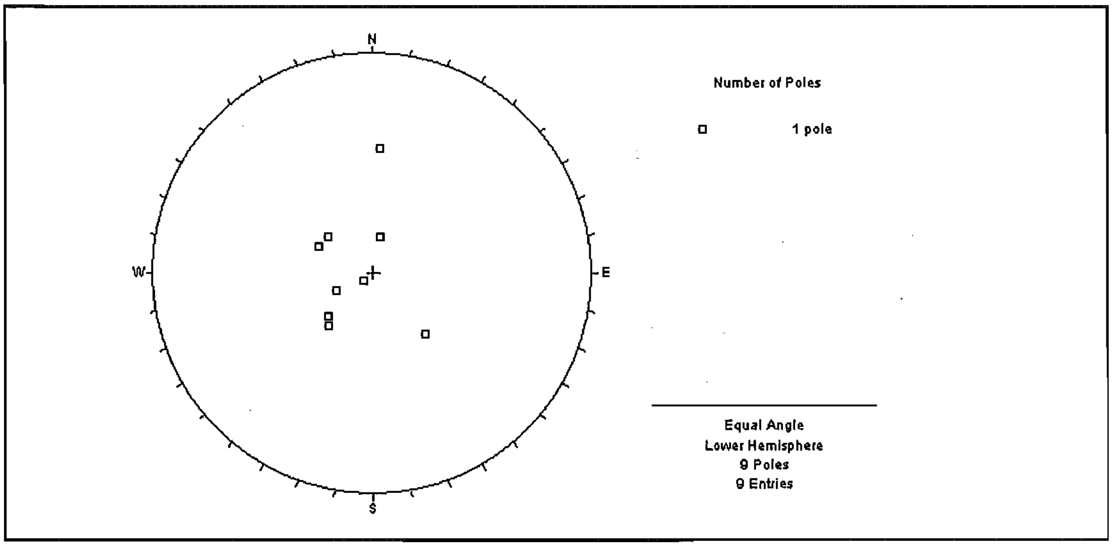

Figure 3.10: Stereographic projection showing the nine poles to defect planes measured in the vicinity of Ella Landslide.

3.1.5 Kate valley field investigation and geotechnical testing summary

Field investigation at the Kate Valley site shows that the site meets requirements for this project as catchment development is being affected by a deep-seated bedding controlled translational slope failure (Ella Landslide) which is inferred to be failing on thin clay rich layer that can be observed within intact stratigraphy and can be termed a "critical stratigraphic horizon". Field observations and scanning electron microscope analysis show that the horizon material is sheared and contains approximately 15% by sample volume of the clay mineral Kaolinite (from X-Ray diffraction analysis), and is a horizon of minimum grainsize contained within a ~300 mm thick package of clay rich material. The horizon is inferred to have a depositional origin, and it is the lithological characteristics that have caused it to become the locus of minor shear displacement, probably during tectonic deformation. The horizon is inferred to have had a controlling influence on the development of the failure surface of the Ella Landslide, and because of the scale of this slope failure it is considered to be very important within the local stratigraphy. Strength testing by ring shear shows that the strength of the critical stratigraphic horizon is relatively low with an effective residual friction angle of 16 – 21°, and an effective residual cohesion of 2.6 – 2.7 kPa.

3.2 Maraetotara Plateau, Southern Hawke's Bay

The study site in Southern Hawke's Bay focuses on several deeply incised catchments developing in Tertiary soft rock terrain that flank the uplifted Maraetotara Plateau which is defined by an erosion resistant sub-horizontal Pliocene limestone cap. Bedding controlled deep-seated landslides are widespread within these catchments (e.g. Pettinga, 1992), and this makes them suitable for analysis in this project as these translational planar block slides and wedge failures are failing on surfaces defined by thin bedding parallel horizons that have a significant impact on catchment development. The Amphitheatre Landslide is a deep-seated retrogressive landslide complex at the head of the Ponui catchment that is representative of a broader population of landslides occurring within the study site.

3.2.1 Field Reconnaissance

Field work in Hawke's Bay was carried out during two trips in February and October 2004, and the Amphitheatre Landslide was confirmed as suitable for this study following two days of field reconnaissance during the first visit. The failure surface of the active landslide appears as an obvious planar and almost horizontal surface (Figure 3.11) and exposures of an in-situ critical stratigraphic horizon on which the Amphitheatre landslide is inferred to be failing occur at several localities in an adjacent catchment.

Figure 3.11: View of the Amphitheatre Landslide at the head of the Ponui catchment. The two near-horizontal, planar failure surfaces of the landslide complex are indicated with arrows and these are mantled by a layer of actively moving landslide debris. Photograph taken looking north east from 6137750N 2843500E (NZMG 260 series map sheet V22).

3.2.2 Field Mapping

The geology of this area has been mapped thoroughly (Pettinga, 1980) and hence mapping for this project focused on geomorphic detail of the Amphitheatre Landslide site and its immediate surrounds. Field mapping and air photo interpretation (NZ Aerial Mapping Ltd runs 1719/72-73 flown 1952, 3833/23-25 flown 1964, 3832/21-26 flown 1964, 3831/21-27 flown 1965, SN5761/25-26 flown 1980 and SN9410/9-10 flown 1995) methods provided adequate data for compilation of a geomorphic map at a scale of 1:10,000 (Figure 3.12). A main focus at this site is to define the stratigraphic levels on which the Amphitheatre Landslide is failing and to correlate this to other deep-seated bedding controlled landslides in this study area. For accurate topographic control a differential GPS survey using Trimble© ProXR GPS units was carried out on the Amphitheatre Landslide and selected landslides is adjacent catchments, using the "W trig" (6139200N 2844100E NZMG 260 series map sheet V22) as the base station for differential correction.

3.2.3 Defect Analysis

Defects have a controlling influence on catchments developing in the Makara Formation rock mass (Pettinga, 1992). Landslide block release, stream incision and hillslope geometry are all subject to some control from rock mass defects. Some defect quantification has been carried out for adjacent landslides (Pettinga, 1980; Pettinga, 1987a, 1987b), however, no data is available in the immediate vicinity of the Amphitheatre Landslide.

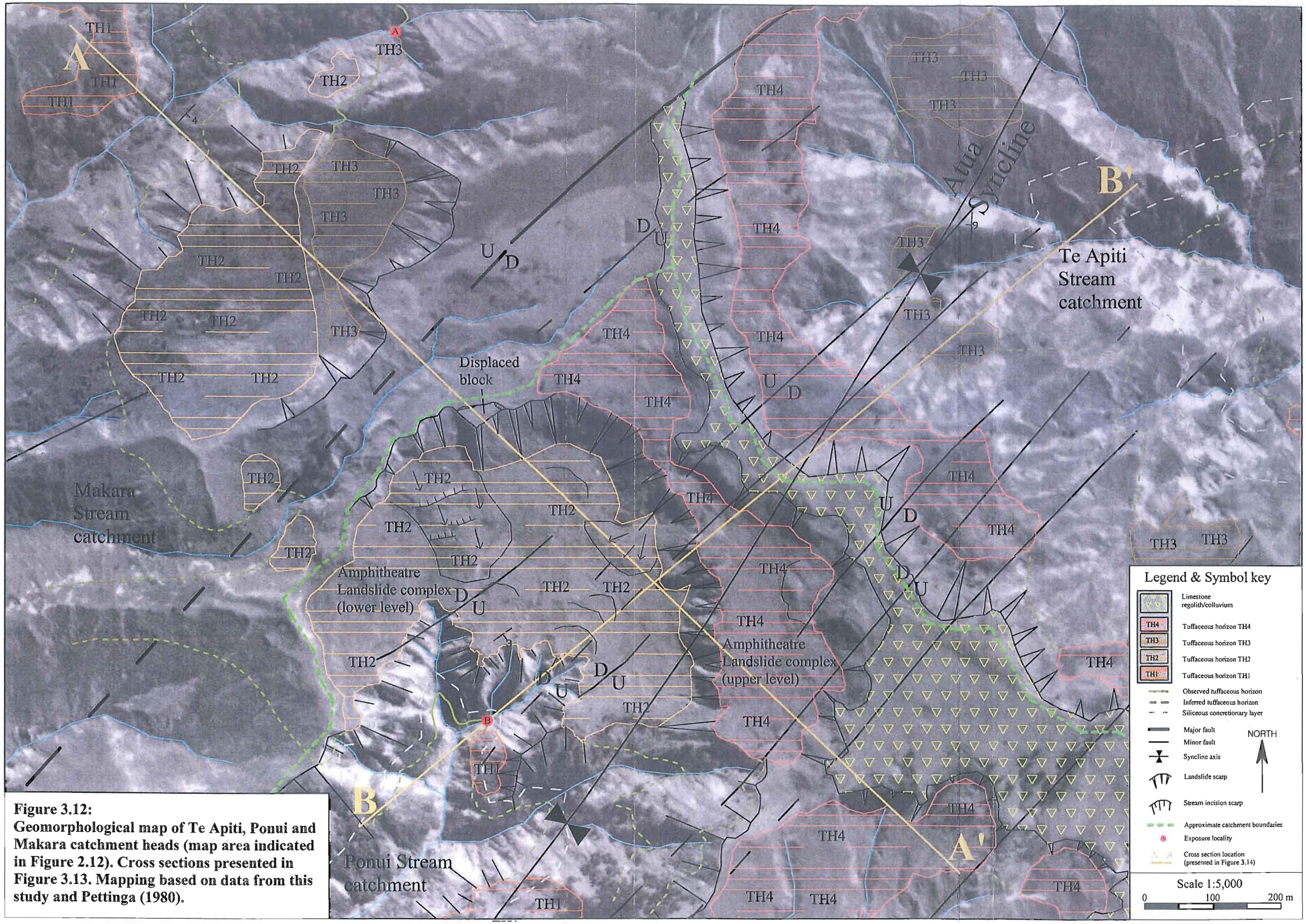

Figure 3.12:
Geomorphological map of Te Apiti, Ponui and Makara catchment heads (map area indicated in Figure 2.12). Cross sections presented in Figure 3.13. Mapping based on data from this study and Pettinga (1980).

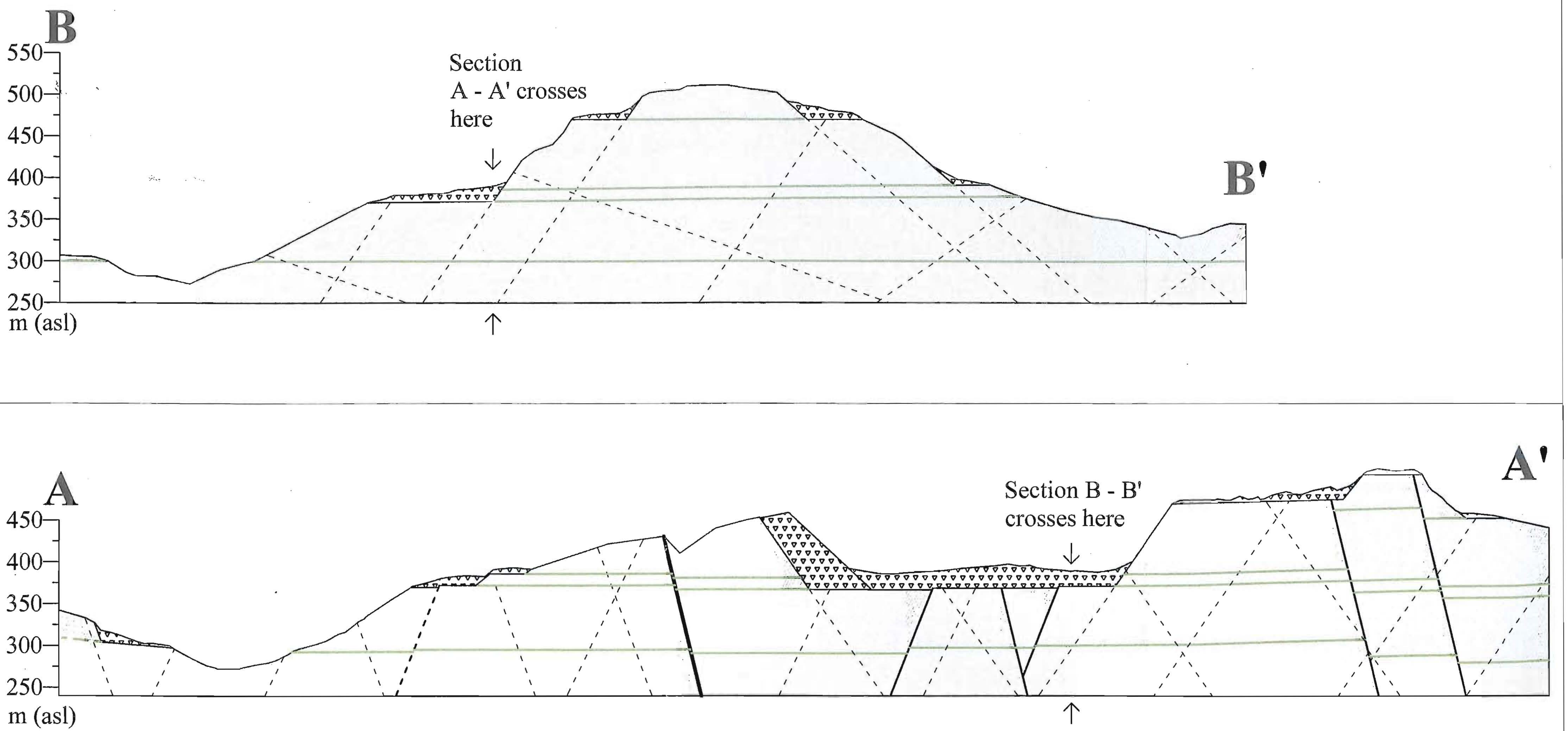

Figure 3.13:
Cross sections A - A' and B - B' illustrating stratigraphic control on landslide location between the Te Apiti, Ponui and Makara catchments (section locations shown in Figure 3.12).

During field mapping the orientations of all defects (faults and joints) encountered were measured. Only major defects were considered (i.e. those with several metres persistence) as many defects are considered to be non-tectonically induced (e.g. by mechanisms such as exfoliation in response to topographically induced stress relief) and hence are not persistent throughout the basin. A contour plot of poles to observed and measured defect planes is shown in Figure 3.14.

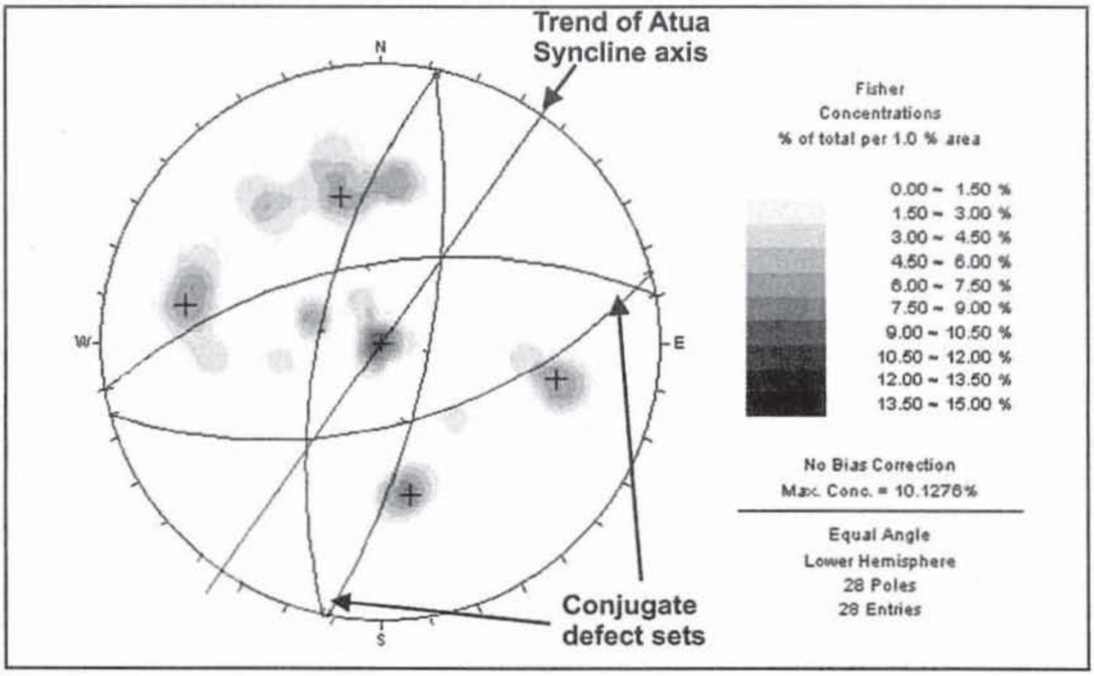

Figure 3.14: Stereoplot showing poles and planes to conjugate defect sets measured in the Makara Formation surrounding the Amphitheatre Landslide, shown in relation to the trend of the Atua Syncline axis.

3.2.4 Critical stratigraphic horizon investigation

Pettinga (1980) first suggested that deep-seated landslides adjacent to the Maraetotara Plateau area are failing on thin bedding parallel tuff layers. Field mapping for this project supports this, with several stratigraphy in-situ tuff beds laterally traceable to landslide failure surfaces, and these tuff beds may hence be considered as critical stratigraphic horizons.

As the origin of the tuff beds within the Makara Formation is documented and reasonably well understood (see Section 2.5.1) it is not considered necessary to characterise them in the same manner as has been done for the unusual and rarely observed critical stratigraphic horizon in the Tokama Siltstone. Cursory microscope examination of the tuff material does show it to be made up of glass shards and pumice fragments, which is in agreement with a tuffaceous origin. The major focus of investigations in terms of critical stratigraphic horizons within the Makara Formation is to make a correlation between landslides which are failing on the same horizon. The aim of this is to determine the extent of critical stratigraphic horizon control on the geographical occurrence of deep-seated landslides within the study area catchments and in the wider region. If tuff layers control where and at what depth deep-seated landslides are occurring, the strength of the tuffaceous material will be the controlling strength used to model landslide stability. Hence it is necessary to obtain samples of the in-situ tuffaceous material for strength testing.

No exposures of in-situ tuffaceous material are available directly adjacent to the basal failure surface of the Amphitheatre Landslide, however, a tuff layer located nearby (shown in Figure 3.15) was initially correlated to the lower basal failure surface of the Amphitheatre Landslide complex. At all locations where exposed this tuff layer is highly oxidized and appears to be partially cemented and in places contains polished surfaces indicative of shear displacement. The layer can be traced laterally for tens of metres which confirms its continuity within the stratigraphy. Subsequent investigation and correlation (based on GPS survey data) determined that this horizon was unlikely to be coincident with the basal failure surface of the lower level of the Amphitheatre Landslide complex but rather occurs ~14 m above it, within stratigraphy between the failure surfaces of the upper and lower levels of the landslide complex. This tuffaceous horizon is correlated with the basal failure surface of several deep seated landslides in the Makara and Te Apiti catchments (landslide correlation is discussed further in Chapter 6). The importance of tuffaceous horizons to deep-seated slope stability is clearly evident from the numerous exposures of tuffaceous horizons of similar character that could be visually correlated to the basal failure surfaces of deep-seated landslides in the Te Apiti, Ponui and Makara catchments. Subsequently the basal failure surface of the lower level of the Amphitheatre Landslide complex is confidently inferred to be defined by a tuffaceous horizon of similar nature to the one shown in Figure 3.15. Further investigation in the incised bed of the Ponui Stream revealed a lower tuffaceous layer occurring some 70 m stratigraphically below the failure surface of the lower level of the Amphitheatre Landslide complex (Figure 3.16).

Figure 3.15: Tuff layer initially inferred to be controlling the failure surface of the Amphitheatre Landslide. The tuff is the horizontal brown layer below the 0.63 m long geopick. Location of exposure indicated as Exposure locality "A" in Figure 3.12.

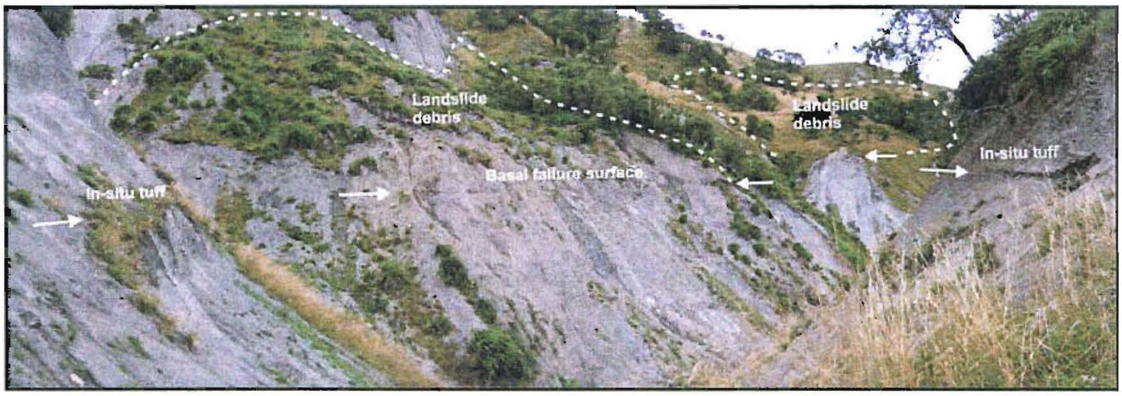

Figure 3.16: Exposure of an in-situ tuffaceous layer (indicated) occurring stratigraphically below the Amphitheatre landslide. Note the lateral continuity and planar nature of the layer and the occurrence of deep-seated slope failure on the horizon. Photograph taken looking south down Ponui catchment from 6137750N 2843700E (NZMG 260 series map sheet V22).

Examination of the lower contact of the tuff showed polishing at the contacts with the siltstone (Figure 3.17) indicating that shear displacement has occurred, however, no sense of shear direction was determined. This tuffaceous horizon can be traced laterally for several tens of metres and, shows signs of shearing in at least two other locations. Two deep-seated slope failures have occurred on this horizon in the immediate vicinity of this exposure (refer Figure 3.16), confirming its status as a critical stratigraphic horizon. During excavation of this critical stratigraphic horizon at one location slope parallel cracks are observed and indicate that some degree of slope relaxation has occurred. The tuffaceous material shows no evidence of oxidization at any of these locations.

Figure 3.17: Surface polishing indicating shear displacement within the tuff layer stratigraphically below the Amphitheatre Landslide complex. Photograph taken in excavation of critical stratigraphic horizon at right of Figure 3.12.

3.2.5 Material strength assessment

Intact block sampling was considered for the purpose of strength testing of the Makara Formation tuff layers, however, the idea was discarded as there is a significant strength contrast between the tuff and the surrounding material which makes intact block excavation impractical. As discussed in Section 3.1.3 the residual strength derived from ring shear testing is considered to adequately define the strength of a critical stratigraphic horizon.

The critical strength controlling the stability of the Amphitheatre Landslide is the residual strength of the tuff layer which defines the stratigraphic location of the failure surface. As other tuff layers observed in exposure are heavily oxidised, which may affect material strength, the lower (non-oxidised) layer was sampled for strength testing. For this it must be assumed that tuff layers throughout the stratigraphy have inherently similar properties and X-Ray diffraction analysis comparing two tuff layers (Figure 3.18) shows that they have the same mineralogy but in different concentrations. It is possible that some mineral contamination has occurred (either syn- or post-depositional) from the surrounding Makara Formation to alter these concentration values.

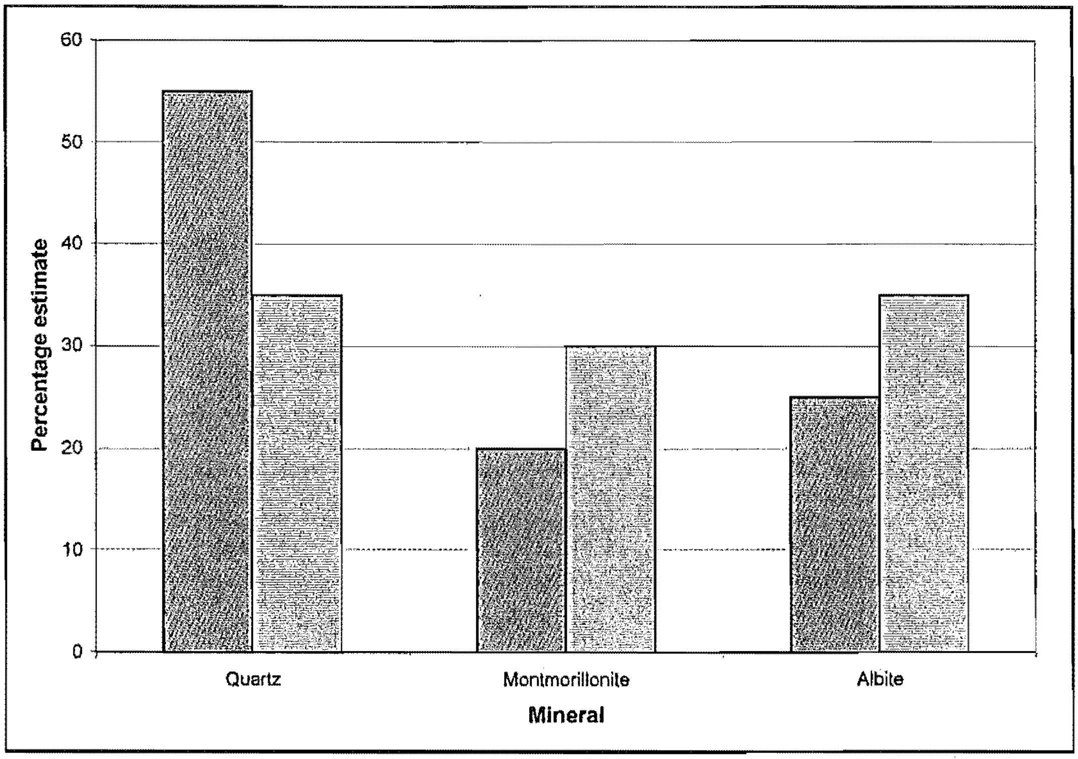

Figure 3.18: X-ray diffraction comparison of two separate tuff beds from the Makara Formation. Samples obtained from exposure locality "A" (light grey) and exposure locality "B" (dark grey) in Figure 3.12.

While the intact strength of the Makara Formation is not tested, from field description it can be described (New Zealand Geomechanics Society, 1988) as being a very weak rock with an unconfined uniaxial compressive strength of 1 – 5 MPa. The critical stratigraphic horizon material, however, would be defined as being a soft soil (undrained compressive strength of 25 - 50 kPa). These descriptions are indicative only, and the Makara Formation strength varies throughout the stratigraphy. Despite the qualitative nature of this strength assessment, it is the strength contrast between the Makara Formation mudstone and the various critical stratigraphic horizons that is particularly significant, in terms of bedding controlled slope stability, and this contrast is always large.

Sampling methodology and preparation

The critical stratigraphic horizon exposure selected for sampling was chosen to be that least affected by weathering. In the stream bed the almost constant water coverage and associated saturation means that little or no oxidization of the tuff has taken place. An approximately 0.3 x 1.0 m area of the lower tuff layer was exposed, directly adjacent to the stream flow, by excavation of the overlying Makara Formation mudstone. This allowed selective sampling of the tuff material, however, numerous sub-angular to sub-rounded pebble size clasts of mudstone were mixed with the tuff and some of these were included in samples. All samples were bagged sealed to preserve moisture content for subsequent laboratory testing.

The critical shear strength for an in-filled defect, where either clasts are matrix supported or defect asperities do not come into contact, is that mobilised in the gouge or matrix (Hoek, 1998). Although the tuffaceous layers in the Makara Formation contain some mudstone clasts, these are supported by the tuffaceous matrix and it is this material which will define the shear strength relevant to a deep-seated slope failure utilizing this horizon as a failure surface. Sample preparation for ring shear testing requires that material fragments > 2.0 mm are removed (British Standards Institution, 1990) and mudstone clasts > 2.0 mm were removed by hand (clasts shown in Figure 3.19).

Figure 3.19: Mudstone clasts removed from Makara Formation tuff prior to ring shear testing.

Ring shear testing results

Three tests were conducted on the selected Makara Formation tuff horizon and these produced very similar results (Figure 3.20). The linear relationship of data points in all test results gives confidence that the derived strength parameters are reasonable and are representative of the tuffaceous critical stratigraphic horizon tested and subsequently of other tuffaceous horizons in the Makara Formation. The strength parameters determined by ring shear testing of this specific critical stratigraphic horizon, and that are representative of the numerous other critical stratigraphic horizons in the Makara Formation, are $\Theta_R' = 2 - 5°$ and $C_R' = 3.8 - 14.2$ kPa.

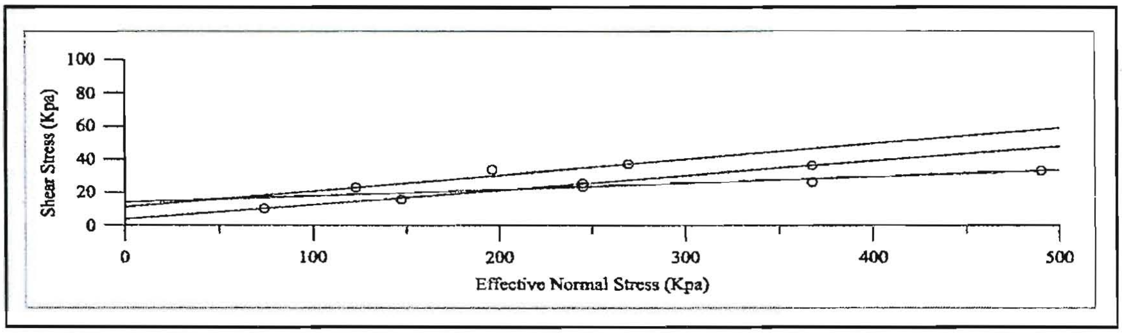

Figure 3.20: Data from the three ring shear tests carried out on the representative Makara Formation tuff horizon.

3.2.6 Durability testing

Justification for testing

Field observations of the Makara Formation show it to be highly prone to degradation by slaking. Blocks of intact mudstone material lying exposed to weathering appeared to degrade rapidly (Figure 3.21). Although an extensive program of durability testing is considered unnecessary for this project, it is interesting to quantify the durability of the Makara Formation.

Figure 3.21: Field degradation (slaking) of intact Makara Formation mudstone near the head scarp of the Amphitheatre Landslide. Outcrop exposure located at 6138200N 2843950E (NZMG 260 series map sheet V22).

There are several durability or slaking laboratory testing methods available for classification of rock behaviour, however, not all are suitable for application to rock containing swelling clays. The Slake Durability Test (see Atkinson et al. (1978) for test procedure) was considered for this study as it is one of the most widely used measures of rock durability. It was found that the material would be classed as being of very high durability ($I_{D2} > 95\%$), however field observation do not support this. The reason for this anomalous material behaviour is inferred to be that oven drying of the sample at 105° C affects the ability of clay minerals to uptake water, hence inhibiting the rock material degradation process.

Drying a sample containing swelling clay minerals to 105° C is clearly not suitable to replicate the material behaviour of interest and hence other methods have been considered. The modified Jar Slake Test (Czerewko and Cripps, 2001) is a more suitable test as it considers the break down of the material without any mechanical action. The specimen is cut to a 45-50 mm sided cube, oven dried at 60° (which avoids the issue of clays mineral structure damage) for 72 hrs and then immersed in water. The sample is periodically observed and its behaviour noted at pre-defined intervals. The test is essentially passive as the sample is not disturbed during the test, though it does not consider the effect of wetting and drying on the specimen.

It is likely that the most suitable test to replicate the physiochemical process thought to cause this behaviour would be one in which the sample is subject to wetting and (accelerated but moderate temperature) drying cycles on the scale of days, i.e. samples are mostly dried out before being immersed in water again. The time scale required for a test such as this would be prohibitively long (weeks to months) but would replicate natural processes in a more realistic manner.

Durability testing results

Modified jar slake results show that the material is "non-durable" or Ij'=7 on an 8 stage scale, where 1 is "no visible sign of specimen deterioration..." and 8 is "total sample disintegration...". This agrees with field observations, where most large pieces of Makara Formation exposed to atmospheric weathering are in the process of, or have broken down to fragments ranging from millimetres to tens on centimetres (Figure 3.21).

3.2.7 Hawke's Bay field investigation and geotechnical testing summary

Detailed topographic information and field mapping confirms that the Amphitheatre landslide and other landslides in adjacent catchments are failing on tuff layers. Several discrete tuff horizons represent critical stratigraphic horizons that can be correlated across catchments and are associated with the development of multiple deep-seated bedding controlled landslides. Testing of representative critical stratigraphic horizons shows them to be very weak with derived strength parameters of $\Theta_R' = 2 - 5°$ and $C_R' = 3.8 - 14.2$ kPa. The disaggregating of the slake prone Makara Formation with wetting and drying cycles is readily observable in the field and the material is classified as non-durable by laboratory testing.

3.3 Field investigation and geotechnical testing summary

Field investigation shows that landslides in both field areas have failure surfaces coincident with thin, weak layers which define critical stratigraphic horizons within the relevant Tertiary soft rock formations. While the origin of the critical stratigraphic horizons in the two field areas differs, they have many similarities. In both cases the horizons are thought to be laterally continuous over significant distances through stratigraphy, meaning that there is potential for more than one landslide to be occurring on the same critical stratigraphic horizons throughout a landscape developing in that particular lithology. In several locations where critical stratigraphic horizons were excavated for sampling, pre-shearing has occurred. This indicates that material is at or near its residual strength and means that any deep-seated slope failure occurring on such a horizon would mobilise the residual strength of this material as oppose to the peak strength of either the critical stratigraphic horizon or the intact soft rock formation. Geotechnical testing shows that the strength of critical stratigraphic horizons is very low with $\Theta_R' = 2 - 5°$ and $C_R' = 3.8 - 14.2$ kPa for the Hawke's Bay site and $\Theta'_R = 16 - 21°$, and $C_R' = 2.6 - 2.7$ kPa for the North Canterbury site. The contrast between the strength of the intact material and the critical horizon is especially significant in terms of where the failure surface of a bedding controlled slope failure is likely to occur and laboratory testing and field observation shows that this contrast is very high in both field sites.

Chapter Four

4.0 Development of representative landslide models

Within each of the two field sites selected for this project (Section 2.3), a representative deep-seated landslide is chosen for more detailed analysis of slope stability under both static and dynamic conditions. The chosen landslides are considered to be representative of a broader population of Holocene aged, deep-seated, bedding controlled landslides occurring in New Zealand Tertiary soft rock terrain, and have clearly defined morphological features relating to the failure mechanism which will allow an assessment of slope stability at the time of failure.

Any quantitative slope stability analysis requires definition of specific parameters including the pre-failure slope geometry, material properties and the mode of initial landslide movement. For the specific type of slope failure under consideration in this study, critical elements of a quantitative slope stability assessment specifically include rock mass defect orientations, the strength of the stratigraphic failure surface horizon and realistic hydrological conditions at the time of failure. In a broader context it is also important to consider other geological and geomorphological components which have allowed the landscape, rock mass and rock material to bring specific slopes to a critical condition for instability.

This chapter defines pre and post slope failure morphology of the chosen landslides, critical parameters that have brought the landscape to a condition of instability and the controls on a specific slope failure. The assessment of slope stability in terms of quantitative static and dynamic slope stability analysis is considered in Chapter 5.

4.1 Ella landslide, North Canterbury

Ella Landslide occurs within the small coastal Kate Stream catchment which is developing within a folded sequence of Tertiary soft rock in the North Canterbury Fold and Fault Belt. The failure is a deep-seated bedrock controlled landslide with its basal shear surface occurring within the Tokama Siltstone Formation, a bedded calcareous siltstone and fine sandstone, while the upper part of the slide mass includes the overlying Greenwood Formation, a massive silty fine sandstone defined by a basal shell pebble conglomerate.

The geomorphic setting is presented in Figure 3.1 showing the Ella Landslide and its geomorphic relationship to the Kate Stream catchment.

4.1.1 Landslide morphology

The Ella Landslide is sited on the north (true left) side of Kate Stream approximately 1.5 km up the catchment from the coast. The location of Ella Landslide marks the transition between the deeply incised stream valley in the lower portion of the catchment and the broad area of fluvial/alluvial valley infill in the mid to upper portion of the catchment, which has formed in response to damming of the creek by landslide debris (see Figure 2.4 and Figure 2.7). The landslide morphology is well preserved and may be divided into three distinct geomorphic domains or zones; the head scarp, main slide block and distal landslide debris (Figure 4.1).

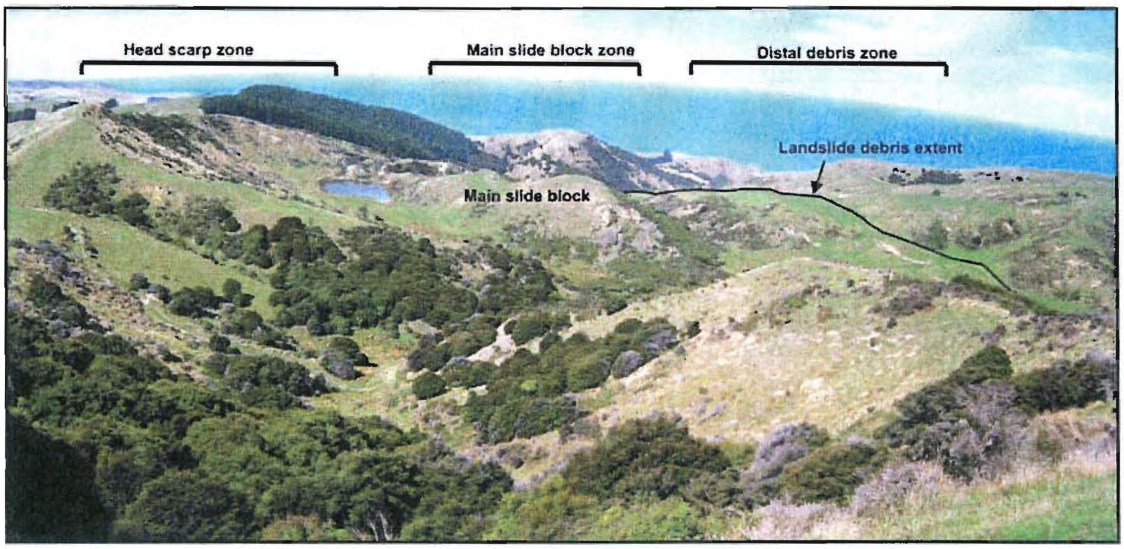

Figure 4.1: Ella Landslide morphology. The pond in the middle distance is formed in the graben between the head scarp and the main slide block (as indicated). Debris has been forced across the valley to the extent defined. Photograph taken looking east from 5790550N 2497750E (NZMG 260 series topographic map sheet N34).

Landslide scarp zone

The semi-degraded landslide head scarp extends to the top of the ridge dividing the catchments of Kate Stream and Dovedale River to the north and strikes approximately E – W. The upper portion of the scarp slopes at approximately 40° to the south and is clear of landslide debris. The lower portion of the scarp is mantled by a colluvial wedge, which thickens as it approaches the main slide block. The left lateral margin of the scarp is well defined and trends approximately perpendicular to the head scarp, however, the scarp forming the right lateral margin is poorly defined as this is covered with landslide debris. Both lateral scarps extend to the crest of minor N-S trending ridges.

Main slide block

The detached main slide block of Ella Landslide defines a graben structure with the head scarp, in which a small lake has formed (Figure 4.1). The slide block forms an irregular hill between 200 and 400 m to the south of the head scarp with steep slopes on the north side of the slide block (graben side) and more gentle slopes to the south, reflecting both the block release surface and the original hillslope morphology. The orientation of the longitudinal axis of the slide block is slightly oblique to the orientation of the head scarp (~15°), and a small graben feature occurs on the crest of the slide block which shows evidence of seasonal pond development.

Distal landslide debris

Debris from Ella Landslide covers a total area of some 400,000 m^2, comprising the main slide block and disintegrated debris covering low hills to the south of the slide block, which consists of both intact blocks and pulverised material from the Tokama and Greenwood Formations. There is some indication of co-failure impact faulting representing toe thrusts within landslide debris material to the south of the main slide block (Figure 4.2), and while close examination of this exposure proved impractical, the location and orientation of these shears strongly indicated their formation in response to the slide block impact. The total extent of landslide debris is shown in Figure 3.1.

4.1.2 Impact of Ella landslide on catchment morphology

The large flat area in the mid-upper part of Kate Stream catchment is defined by extensive alluvial accumulation following the Ella slope failure (Figure 2.7). Landslide debris caused a dam in Kate Stream which became a barrier to sediment removal by fluvial transport, and sediment subsequently accumulated to form a flood plain. Distal landslide debris also dammed a small tributary of Kate Stream, causing accumulation of a minor amount of alluvial material (evident in Figure 3.1). The main valley fill deposit is inferred to be 60 m thick from interpolation of the surveyed longitudinal profile of Kate Stream (Geotech Consulting Ltd, 2002), and ^{14}C dating of a sample retrieved from 4.8 m below the surface of the fill indicates a minimum age of 1461 ±60 yrs BP.

Figure 4.2: Apparent co-failure impact thrust fault development inferred to be the result of block slide impact into debris. Exposure locality "C" in Figure 3.1.

The morphology of the mid-upper catchment is significantly different in the present day than it was prior to the occurrence of Ella Landslide. The catchment morphology at the time of the landslide has a significant bearing on the Ella Landslide slope stability model with respect to the extent and geometry of the pre-failure slope, and must be inferred from the current landslide debris distribution and the projected valley slope geometry.

Incision of the ancestral Kate Stream

Based on the deeply incised nature of Kate Stream below the Ella Landslide, and the cross valley profile upstream of the valley fill alluvial plain, the pre-failure slide mass had become unsupported by the deep valley incision in post-glacial times (early Holocene). Ella Landslide is inferred to have occurred at least 5,000 – 7,000 years ago by Geotech Consulting Ltd (2002) based on accumulation rates of valley fill, defined by a sequence of ^{14}C dates from samples retrieved in an excavated pit in valley fill sediments. The deep incision of Kate Stream must have progressed upstream of the site of Ella Landslide, and possibly as far as the older Cass Landslide (see Figure 3.1), before the Ella slope failed. Mapping indicates that the location of the ancestral channel of Kate Stream at the time of failure was coincident with the main slide block (Figure 3.1). The orientation and profiles of the valley sides upstream of Ella Landslide indicate that the stream axis was likely to have included a meander bend at approximately the location of the Landslide. Upstream of Ella landslide Kate Stream flows in a SW – NE direction, whereas downstream of Ella Landslide the stream turns almost ninety degrees to flow in a NW – SE direction. The location and geometry of the pre-failure slope on the outside of this inferred meander would have probably lead to the development of a steep, possibly undercut slope which may have enhanced the instability of this particular slope.

The depth of the incision at the landslide location can be estimated using the gradient of Kate Stream where it is incised into bedrock. The lower Kate Stream gradient (between Ella Landslide debris and the sea) is approximately 1 in 40 (Geotech Consulting Ltd, 2002) and this is considered to be appropriate for predicting the depth of the stream incision at the location of the Ella slope failure a few hundred metres upstream. This is relevant to the destabilisation of the slope as it is inferred that slope failure would occur once the stratigraphic horizon which defines the landslide failure surface has become exposed.

4.1.3 Rock mass defect control on Ella slope failure

Rock mass defects (including bedding, joints and faults) have a controlling influence on the geometry and failure mode of deep-seated landslides in Tertiary soft rock terrain. At the site of the Ella landslide stratigraphic dips of 15° to the south are coincident with the direction of slope failure.

Ella Landslide failure surface

It is inferred that the Ella Landslide has failed on a single, discrete ~5 mm thick clay rich bedding parallel horizon (refer Section 3.1.2), and the stratigraphically confined and lithologically discrete Kaolinite-rich clay horizon is inferred to have a depositional origin. The probable low energy environment in which the formation was deposited implies a sediment cloud of clay-size material settling out of the water column and blanketing a geographically extensive area of the sea floor. This may or may not have been preserved as a laterally continuous layer throughout the entire extent of the Tokama Siltstone, but it is inferred to be continuous at least across the failure surface of Ella Landslide, and to the exposure observed at the downstream limit of the failure debris (exposure locality "A" in Figure 3.1). The failure surface of the Ella Landslide is correlated from this key outcrop exposure to the landslide failure area using the elevation of the horizon as defined by GPS and projecting it onto a cross section perpendicular to bedding. While the simple structure in the study area and the close proximity of the exposure to the landslide source area gives confidence in the correlation, there is a degree of (unquantified) associated error. If the critical stratigraphic horizon observed in exposure does not define the failure surface of the

Ella Landslide, it is inferred with significant confidence that a horizon with very similar geotechnical properties would define the basal shear plane. The use of structure contours to project the stratigraphic horizon through topography to the landslide was considered to be of limited value due to the lack of accurate topographic and limited bedding orientation data.

Shearing observed in the clay rich layer is inferred to be post-depositional, and the preferred mechanism for this is minor bedding parallel shear during flexural slip folding of the nearby Kate Anticline - Teviotdale Syncline pair (discussed in Chapter 1). A kinematic indicator of the direction of slip on the critical stratigraphic horizon (Figure 3.6) suggests that shear has occurred in the direction of the trend of the synclinal axis. For a generalised flexural slip shear surface, displacement would be expected to be in the dip direction, however, in this case the shear orientation in the fold axis direction may be related to contraction due to the termination of the Teviotdale Syncline against the Kate Anticline.

Landslide release mechanism

Many bedding controlled deep-seated soft rock landslides are documented to utilise rock mass joint sets as both head and lateral release mechanisms (Thompson, 1981; Pettinga, 1987a, 1987b; Bell and Pettinga, 1988). The lithological formations in which the Ella Landslide has occurred have very few persistent defects in the vicinity of the landslide (see Section 3.1.4), however, it is considered possible that the slide mass has released on some form of rock mass defect surface (most likely a joint surface).

Figure 4.3 shows typical tectonic joint sets developed in a low amplitude fold, indicating the orientation of joints in stereographic projection at different locations throughout a syncline/anticline in relation to the trend of the fold axis (marked by the b-axis). The location of the Ella Landslide within the Teviotdale Syncline is analogous to the situation in stereographic net "C", which shows joints striking parallel to the trend of the fold axis and dipping oblique to bedding. The head scarp of Ella Landslide trends approximately parallel to the axis of the Teviotdale Syncline and is oriented at a steeply oblique angle to bedding. It is inferred from the orientation of the head scarp, and the mechanism and nature of tectonic joint development in Figure 4.3, that it is possible for the Ella Landslide to have released on such rock mass defect surfaces.

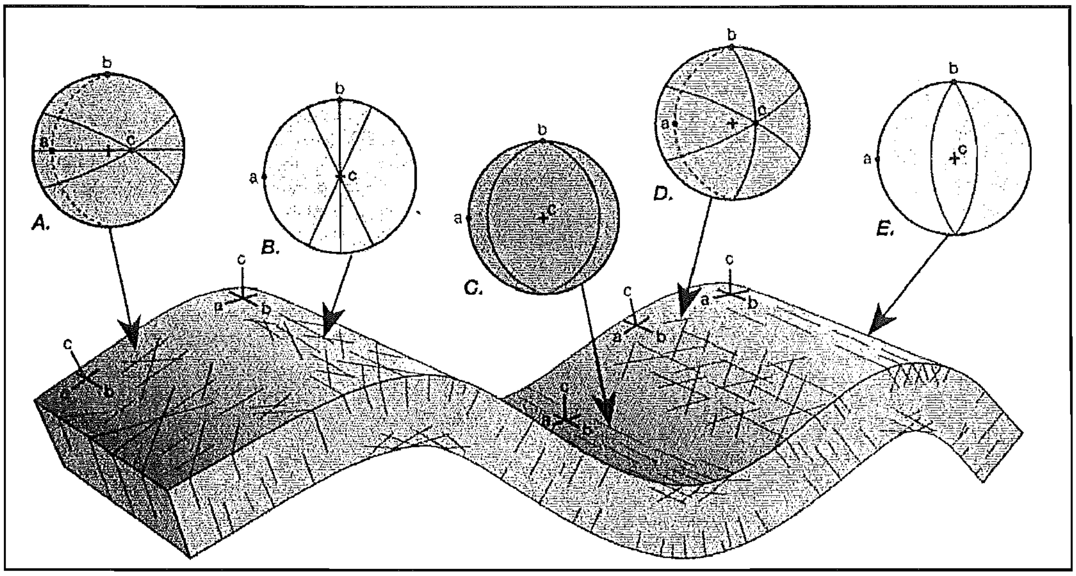

Figure 4.3: Schematic representation of the development of tectonic joints due to low amplitude folding. Lower hemisphere stereographic projections show typical joint orientations at locations within the folding in terms of three principle axis orientations. From Twiss and Moores (1992).

4.1.4 Hydrogeological conditions

Hydrogeological considerations in terms of slope stability analysis include the elevation of the free water table with respect to the landslide failure surface, and the presence of any confined and/or perched water tables. Hydrogeological conditions are always going to be a problematic aspect of slope failure analysis for a pre-historic landslide, however, inference about hydrogeology can be made based on data from present day conditions. In this respect it is fortunate for this study that there is considerable data available from the investigation programme for the new regional landfill in Kate Valley.

The free water table in parts of upper Kate valley has been defined based on borehole monitoring (Geotech Consulting Ltd, 2002) which shows that (away from water courses) the water table may rise a small amount with topography, but that this variation is likely to be minimal and the groundwater has a mostly sub-horizontal profile beneath ridges between valleys. Unconsolidated sand beds within the Tokama Siltstone Formation may act as seepage paths (Figure 4.4) as they are likely to have a higher permeability than the surrounding siltstone. These beds may act as confined aquifers and Geotech Consulting Ltd. (2002) infer that elevated artesian pressure within beds of this nature is a likely mechanism for initiating deep-seated bedding parallel slope failure in the area.

Figure 4.4: Unconsolidated sand bed acting as a seepage path within the Tokama Siltstone (dark layer indicated with arrow). Location of exposure shown in Figure 3.1 xx needs grid reference.

4.1.5 Pre-failure slope model

The Ella Landslide slope failure model is based primarily on air photo interpretation in conjunction with GPS survey data. GPS topographic mapping has allowed detailed delineation of topographic sections through scarp and debris zones and along in-situ bedrock ridges formed adjacent to the lateral scarps, and these are assumed to represent pre-failure valley-side topography. The original slope indicated in Figure 4.5 has been inferred from a combination of the lateral scarp ridge profile and the projection of the slide block up-dip along the failure plane, and shows the projected pre-failure topography, as well as the current topographic configuration of the landslide debris. It is the projection of the slide block to the head scarp that suggests that the original ridge crest was at a higher elevation than occurs today.

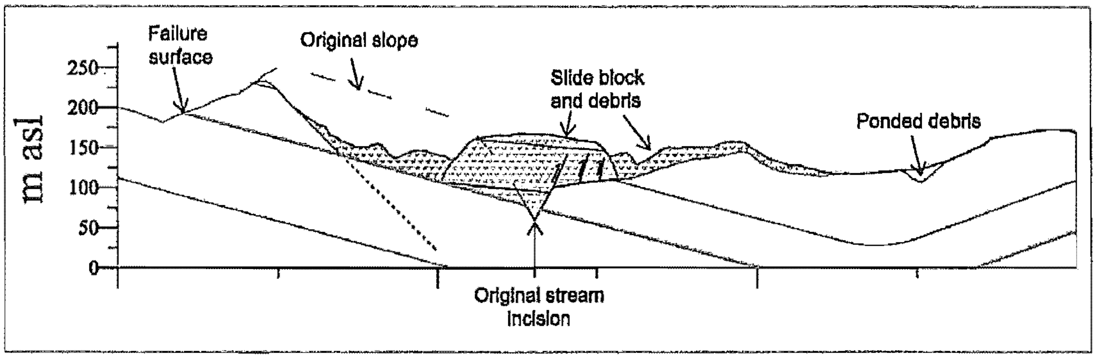

Figure 4.5: Representative cross section of the Ella Landslide showing both the original (pre-failure) slope geometry and the current, post-failure geometry. This representative cross section is based on the cross section presented in Figure 3.1.

4.1.6 Landslide failure model

The final failure of the Ella Landslide is inferred to have been rapid, based on the distance that toe debris has been emplaced across the valley (Figure 3.1). The triggering mechanism for the slope failure is considered in detail in the following chapter and the mode of debris emplacement is considered here. Two models are proposed for slope failure and debris run-out which could have resulted in the current debris configuration:

- The laterally unsupported slide mass fails in one motion, with the main slide block coming to rest in the incised Kate Stream bed. As this motion is rapid material is forced past the Kate Stream valley and this results in a mantle of debris covering the opposing slopes; or

- When the main slide block fails there is already a quantity of debris sitting on the failure surface, related to a previous slope failure. When the slide block travels rapidly down dip it collects debris on the landslide shear surface and impacts on the earlier emplaced slide mass in the valley, forcing it onto and across the opposing slope resulting in a mantle of debris covering these slopes.

It is apparent from field mapping and vertical air photo interpretation that there may have been some deviation for the model of a purely translational, planar block slide failure. Plan view anticlockwise rotation and back rotation of the main slide block is inferred from both the strike of the slide block being oblique to the orientation of the head scarp (refer Figure 3.1), and from the dip of the pebble shell conglomerate at the western end within the slide mass. A three point analysis on the shell pebble conglomerate (based on GPS data points) indicates that it dips south at 4° as oppose to the 15° dip of the intact stratigraphy. Back rotation is inferred to have occurred in the last stages of debris emplacement as the slide block came to rest in the incised stream bed. Plan view rotation may be due to several factors including, failure not occurring directly down dip, the occurrence of higher shear resistance on one lateral scarp; or, limited space available to accommodate the slide block at one side of the failure. As the slide mass is inferred to have slid into a meander bend in the ancestral Kate Stream, it is likely that the space available to accommodate the slide mass would be variable along the longitudinal stream profile, and slide block rotation would occur due to differential impact.

While the mode and configuration of landslide debris emplacement has little bearing on slope stability modelling, it has a significant influence on the geomorphic development of the Kate Stream catchment, and in Chapter 6 catchment evolution is considered in terms of the impact of Ella Landslide. The inferred pre-failure cross section, which has been developed based on data discussed in this chapter (Figure 4.5), can be used to model the stability of the Ella Landslide at the moment of failure and this is considered in Chapter 5.

4.2 Amphitheatre landslide, Southern Hawke's Bay

The Amphitheatre Landslide is a retrogressive planar block slide complex, failing on stratigraphically controlled surfaces at the head of the small-moderate sized coastal Ponui Catchment which is developing adjacent to the southeastern edge of the Maraetotara Plateau in Southern Hawke's Bay. The landslide has previously been documented by Pettinga (1980; 1992) and is considered to be representative of the low angle retrogressive planar block slides in the three catchments under consideration in this study. The Amphitheatre Landslide is failing in the Makara Formation flysch sequence of alternating finely bedded and graded, calcareous sandstone, siltstone and mudstone units, and occurs on two separate basal failure planes which split the complex into two separate levels (Figure 4.6), which define bench-like morphological features in the landscape.

Based on the less degraded appearance of the lower level of the Amphitheatre Landslide complex, it is clearly this level which is predominantly active in the current environment. Lower level retrogression occasionally induces failure to occur on the upper level (moderate failure of both levels in 1974 was documented by Pettinga, 1980), however, it is the activity of the lower level which is of interest for modelling the stability of the Amphitheatre Landslide as representative of active landslide complexes within the study area. To model the stability of this landslide complex it is necessary to define the evolution of the landslide in terms of large scale landscape development (e.g. long-term tectonic and climatic forcing), and controls at the individual slope scale (e.g. critical material strengths and short-term tectonic and climatic forcing) so that the failure mode critical to ongoing slope instability can be assessed using numerical slope stability modelling methods.

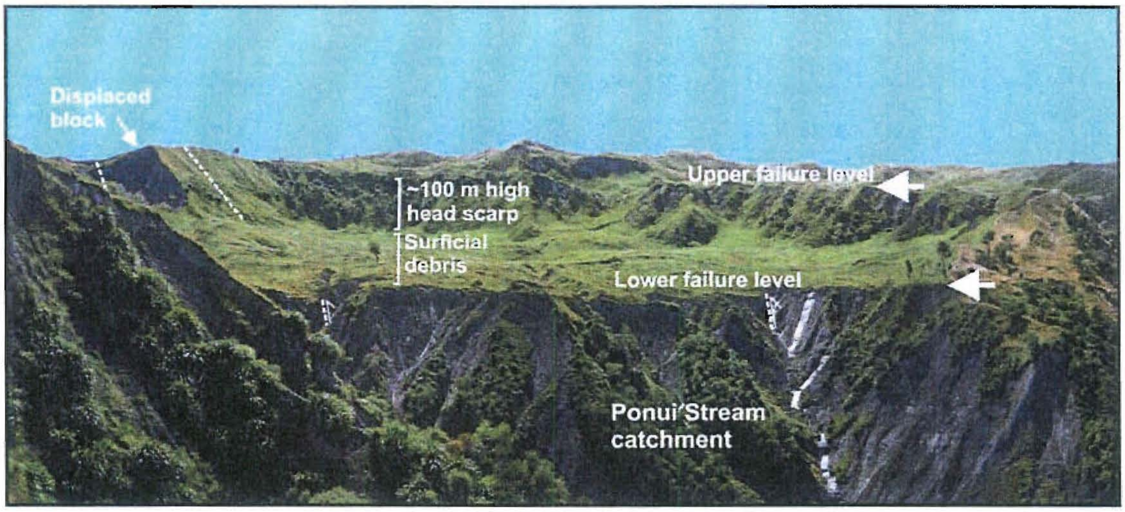

Figure 4.6: View of the Amphitheatre Landslide from Ponui catchment showing the two failure levels, the lower level scarp, surficial debris and the incised stream valley below. Displacement of the lower level failure surface by faulting is indicated (dashed lines in streams) and the extent of a block which has displaced from the scarp is shown at the left of the photograph. Photograph taken looking north east from 6137750N 2843500E (NZMG 260 series map sheet V22).

4.2.1 Landslide morphology

The Amphitheatre Landslide complex can be considered in terms of three major geomorphic domains or zones; the degraded upper level, the active lower level head scarp, and lower level surficial debris mantling the failure surface (Figure 4.6). Associated features such as the deeply incised Ponui Stream gully below the landslide complex have a direct influence on landslide activity and evolution, and this fluvial system has removed the majority of debris produced by the landslide complex during its development. The influence of stream incision and other factors on landslide evolution will be considered in the following sections of this chapter.

Degraded upper level

The upper level of the Amphitheatre landslide covers an area of some 85,000 m^2 and is characterised by a smooth hummocky topography on the bench surface and (50-60 m) high scarp, indicative of quasi-stability. Although occasionally involved in slope failures propagating up from the lower level, this part of the landslide complex is considered to be essentially inactive. Figure 4.7 shows the delineated extent of the upper level of the

Amphitheatre Landslide as it occurs today, and indicates the inferred extent of the landslide complex at the time of initiation. The inferred extent of the (ancestral) landslide complex is considered conservative, and it may have covered a significantly larger area in the past based on the inferred extensive lateral continuity of the critical stratigraphic horizon which forms the failure surface.

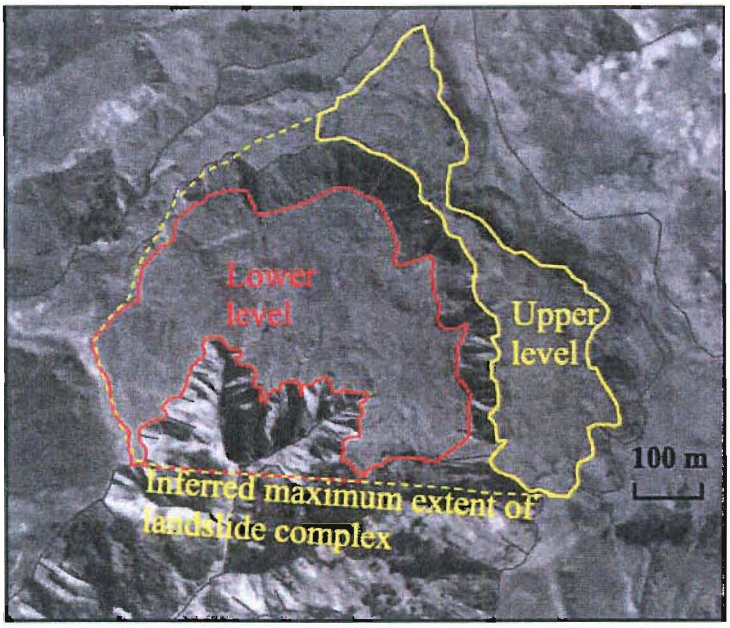

Figure 4.7: Map view of the Amphitheatre Landslide showing the aerial extent of the upper and lower levels as they occur today and as inferred extent of both at the time of landslide initiation. Map based on NZ Aerial Mapping vertical air photo 3832/24 flown 1964, north is directly up the page.

Lower level head scarp

The entire scarp for the lower level of the landslide is defined as a head scarp as there is no clear definition of lateral scarps. The scarp is approximately 100 m high and is steep and angular for the north and eastern portion as it is predominantly formed of intact Makara Formation mudstone with a debris mantle only on the lower slope. The western portion of the scarp is less well defined as this is predominantly covered in landslide debris.

At the left side of the landslide looking from Ponui Catchment an approximately 3.2×10^5 m^3 block of intact Makara Formation Mudstone has detached from the scarp and is displaced by a small amount (Figure 4.8). On the scarp to the south west of this displaced block there is debris at the base of the scarp consistent with the full degradation and failure of a similar block.

Figure 4.8: Slightly displaced block of intact Makara Formation mudstone on the head scarp of the lower level of the Amphitheatre Landslide, note the tension crack which defines the extent of the top surface of the block next to person for scale. The location of the block is indicated in Figure 4.6.

Lower level surficial debris mantled failure surface

The lower level of Amphitheatre Landslide defines a relatively flat semi-circular area of approximately 145,000 m^2 which is defined by the removal of intact bedrock on a discrete basal failure surface (arrowed in Figure 4.6), which is mantled by 1 – 20 m of degraded landslide debris. This comprises degraded mudstone which has become a weak, saturated debris involved in a creeping debris flow that is actively transporting material into the incised Ponui Catchment. This movement is clearly reflected in tension crack features on the debris surface (Figure 4.9) and the activity of the debris flow is partly seasonally controlled (Pettinga, 1980), however, field observation shows it to be constantly active.

Volume estimation of removed landslide debris

The morphologic form of the Amphitheatre Landslide reflects a "deflated" slide mass, in the sense that the bulk of the landslide debris has been removed and transported down Ponui Stream via the creeping debris flow, and Figure 4.7 shows the inferred maximum extent of both the upper and lower levels of the landslide. This inferred extent can be used to calculate a crude volume estimate for material that has been removed from the landslide complex, as defined by the area of the failure surface and scarp height for each level. The lower level, in its current state, is missing an estimated 13 x 10^6 m^3 of material, and the upper level an estimated 4 x 10^6 m^3. If the inferred maximum extent of the landslide complex is considered, then the combined volume of material removed for the upper and lower levels may be in the order of 65 x 10^6 m^3. As mentioned this is considered to be a conservative estimate and only approximates the quantity of material this landslide complex has introduced to the Ponui Catchment sediment flux during the late Pleistocene and Holocene. The calculation includes the material involved in the debris flow on the basal failure surface and takes no account of material bulking. A detailed and accurately quantified analysis of the sediment input from deep-seated landslides to the sediment flux in the study catchments is beyond the scope of this project.

Figure 4.9: Lower level of the Amphitheatre Landslide with arcuate tension cracks reflecting the actively creeping debris on the almost exhumed basal failure surface (movement right to left). The mound of debris at centre right of the photo relates to the degradation of a block recently detached from the head scarp. Photograph taken looking south west from 6138200N 2843950E (NZMG 260 series map sheet V22).

4.2.2 Factors influencing deep-seated landslide occurrence

Certain factors that critically influence deep-seated slope stability, such as rock mass and rock material properties may be considered as latent in the landscape. These factors influence the geometry and spatial distribution of deep-seated landslides but for slope failure to initiate these must occur in combination with temporally variable factors such as catchment incision and short term tectonic and climatic forcing.

Deep-seated landslides in the catchments adjacent the Maraetotara Plateau initiate when stream incision exposes critical stratigraphic horizons, and hence the first occurrence of a landslide will likely be when the failure plane is at stream bed level. As stream incision is driven deeper these early stage landslides progressively become perched at higher elevation in the landscape and persist as active complexes on a scale of $10^2 - 10^4$ years. The Amphitheatre Landslide is perched some 100 m above present day base level and persists to be an active landslide complex, despite the fact that it probably initiated in the late Pleistocene (Pettinga, 1992). The general model for the development and persistence of landslide complexes in this study area is discussed further in Chapter 6.

Catchment incision mechanism and control on slope stability

A key factor that has allowed the Amphitheatre Landslide to initiate and persist to be active is the incised stream network. Stream incision facilitates slope failure by the removal of lateral support of potential slide masses and by exposing weak layers in the stratigraphy which act as failure planes. The Amphitheatre Landslide is inferred to have initiated when the Ponui Stream exposed the upper failure surface at stream bed level (the failure surface is now some 200 m above stream base level) and allowed an initial slide mass to become unsupported.

The incremental lowering of stream base level is directly related to the development of deep-seated bedding controlled landslides at successively lower levels. Rapid base level lowering episodes, driven by long-term tectonic and climatic forcing, lead to periods of accelerated stream incision and gully enlargement and the subsequent exposure of successively lower critical stratigraphic horizons. These critical stratigraphic horizons define the failure surfaces for deep-seated landslides which have a critical role in the evolution of these catchments, particularly at the catchment head (this role will be discussed in detail in Chapter 6) and lead to a pervasive bench-like morphology in the landscape.

The bench-like morphology which results from the widespread occurrence of bedding controlled deep-seated slope failure contrasts with the steep valley walls of deeply incised stream catchments. Periods of accelerated stream incision (driven by rapid base level lowering events) migrates through the catchment from sea-level to catchment head in over steepened reaches or knickpoints. As these knickpoints migrate up the catchment, and the stream bed is incised to a lower elevation, the steep gully slopes must widen to accommodate this.

Mechanism of stream incision and valley side retreat

Incision rates are often directly related to stream power (e.g. Siedl and Dietrich, 1992), however, physical laboratory modelling shows that fine grained sediment does little work in terms of bedrock abrasion (Sklar and Dietrich, 2001). The ephemeral nature of streams in the upper parts of these catchments, and the very fine grainsize of the majority of entrained material, means that the stream power relationship of bedrock incision is unlikely to apply, and mechanical abrasion is not likely to be the primary mechanism of stream incision into bedrock.

A critical mechanism allowing these processes to occur is considered to be a combination of the degradation of the intact Makara Formation mudstone and the presence of intersecting conjugate defect sets. The rock material can be classified in the laboratory as "non-durable" and is clearly highly prone to slake degradation in the field and the rock mass has clearly defined conjugate defect sets (see Chapter 3).

The geographic location of stream networks is controlled by the intersection of conjugate defect sets (Pettinga, 1980), where a comparatively easily eroded zone of rock mass weakness occurs, defining a preferential location for stream incision. Material slaking, related to wetting and drying cycles, results in the formation of a slope parallel layer of surficial debris on intact mudstone in the steep valley walls (Figure 4.10). This material eventually builds up, detaches and accumulates at the slope base to be removed during high intensity precipitation events.

Figure 4.10: Stream incision in the degradation prone Makara Formation. Surficial slaked material shows as dark grey on gully sides and is stripped in the stream bed (light grey). The location of the stream bed is controlled by rock mass joints as visible at the right of the stream bed in the centre of the photo. Photo taken looking west up a tributary to Ponui Stream from 6137850N 2843600E (NZMG 260 series map sheet V22).

As second, third and fourth order gullies propagate from the first order incised channels prominent spurs are formed between the steeply incised gullies. It is apparent that these spurs are then prone to failure on approximately slope parallel shear surfaces (Figure 4.11). It is inferred that a combination of material degradation by slaking in incised gullies and spur failure are two of the primary mechanisms of slope parallel retreat of steep valley walls controlled by base level lowering rates.

Figure 4.11: Slope parallel spur failure in the Ponui catchment involving the Amphitheatre Landslide basal shear surface, note the displaced failure surface contact with topsoil (arrowed). Photograph taken looking west from 6137950N 2843800E (NZMG 260 series map sheet V22).

The overall effect of deepening stream incision into bedrock and stream gully enlargement, by whatever mechanism, is that increasing amounts of the stratigraphic column are exposed including critical stratigraphic horizons that act as failure surfaces for deep-seated slope failures.

Landslide failure surface

The failure surfaces of the Amphitheatre Landslide complex, as well as many other deep-seated landslides in the study area catchments, are coincident with the failure surface of bedding parallel tuff layers (Pettinga, 1980, 1992), and the origin of these tuff beds is discussed in Section 2.5. Minor shear displacement on the intact tuff layer, stratigraphically below the lower Amphitheatre Landslide (Figure 3.17), is inferred to occur in other tuff layers acting as landslide failure surfaces. Shear development reduces the unconsolidated tuff material to at or near its residual strength, enhancing the strength contrast which already exists with the surrounding mudstone lithology. The mechanism of shear displacement in the intact Makara Formation rock mass is considered in terms of the mechanisms discussed in Chapter 1, flexural slip and progressive failure, and a proposed mechanism related to rock mass adjustment due to tectonic deformation.

Flexural Slip

Stratigraphically discrete bedding parallel flexural slip can occur in response to the early onset of folding. Slip is most likely to occur at weak points in the stratigraphy, and the most favourable situation for shear development is when thick competent beds are separated by thin weak beds (Hutchinson and Anonymous, 1995). The tuffaceous layers within the Makara Formation define weak horizons within the stratigraphy as they are completely uncemented and have a significant grainsize and strength contrast with the more competent and thicker mudstone beds of the surrounding Makara Formation. It is likely that the wide (50 – 100 m) spacing of the tuffaceous beds within the Makara Formation enhances the magnitude of shear which occurs within them. Thin and closely spaced weak layers will distribute shear through the stratigraphy, whereas widely spaced thin weak layers will enhance the magnitude of shear development on discrete stratigraphic horizons. Leith (2003) showed that tectonic deformation inducing dips of 13° in a 1 km thick sediment pile would require 22.6% strain within the succession, considered to be sufficient to produce flexural slip shear planes. The dips and sediment thickness in this example are not dissimilar from those in the Makara Formation, and it is inferred that deformation of this succession is sufficient to produce flexural shear given the occurrence of widely spaced weak tuffaceous horizons.

Progressive Failure

In geotechnical terms the Makara Formation is effectively an overconsolidated mudstone and as such may contain some recoverable strain energy in the horizontal direction (discussed in Chapter 1) which may induce shear within the sedimentary succession as this strain is recovered. With the contrasting strength of tuff layers it is possible that preferential progressive failure could induce shear in these layers as the rock mass becomes laterally unsupported due to valley incision. While this mechanism seems reasonable in the near flat lying stratigraphy in the synclinal axis it does not adequately explain shear inferred to occur in the steeper dipping limbs of folds.

Rock mass adjustment

Another possibility for shear development in the tuffaceous horizons in the Makara Formation relates to rock mass adjustment driven by the large scale thrust structures which bound the Makara Basin (refer Figure 2.12), (J Pettinga pers. comm. 2005). Northwest dipping thrust faults propagating through the underlying basement rock will define a wide zone of deformation as they near the surface. Sub-surface fault plane rupture leading to distributed strain and deformation of the cover sequence will result in localised stress concentrations in the discontinuous rock mass (as defined by minor faults, joint sets and weak stratigraphic horizons). To allow deformation of the slab-like cover rock succession the rock mass may adjust by shear displacement on existing discontinuities, and as tuffaceous beds in the Makara Formation define weak stratigraphic horizons they would be a likely focus of this associated shear.

It is likely that all these mechanism are occurring to a certain degree in the Makara Formation rock mass. The likelihood that progressive failure is causing widespread but minor shear displacement on stratigraphic horizons in this folded sequence is considered unlikely and inherited stresses are more likely to develop minor slope parallel slab failures (Prebble, 1992), though these are less pervasive in this lithology than they are in other areas of New Zealand soft rock terrain (e.g. Taihape Mangaweka area, Thompson, 1981). The preferred mechanism for shear development in tuffaceous horizons is a combination of flexural slip at the initial stages of folding and rock mass adjustment in response to thrust fault induced strain distribution and deformation of the cover rock succession.

Rock mass defects

The Makara Formation has well defined rock mass joint sets (refer Chapter 3), and conjugate joint sets in the Hawke's Bay study area have a tectonic origin (see Figure 4.3), (Pettinga, 1980, 1992).

Defect sets form a critical component of landslide development as they control where stream incision occurs (as indicated in Figure 4.10), provide landslide block release surfaces, and subsequently define the geometry of both landslide scarps and hillslopes. Figure 4.12 depicts the situation in the Amphitheatre Landslide where defects directly control scarp geometry. Those joints which were able to be excavated within an intact sample showed no discernable tensile strength (in terms of qualitative field assessment), thus providing an excellent slide block release mechanism, and the overall scarp and slope geometry directly reflects the orientation of conjugate defect sets.

Joints observed in the field are frequently tight, and while some show minor infill this is often found to be a surficial feature related to material weathering. Some rock mass joints have been disturbed and are dilated to the point where they are open and this is inferred to result from rock mass disturbance due to factors such as seismic disturbance and accompanying slope relaxation. The joints observed in the area of the displaced block in the Amphitheatre landslide complex (Figure 4.8) are often open (5 – 20 mm aperture) and this rock mass disturbance is likely to relate to the failure of the displaced block (Figure 4.8).

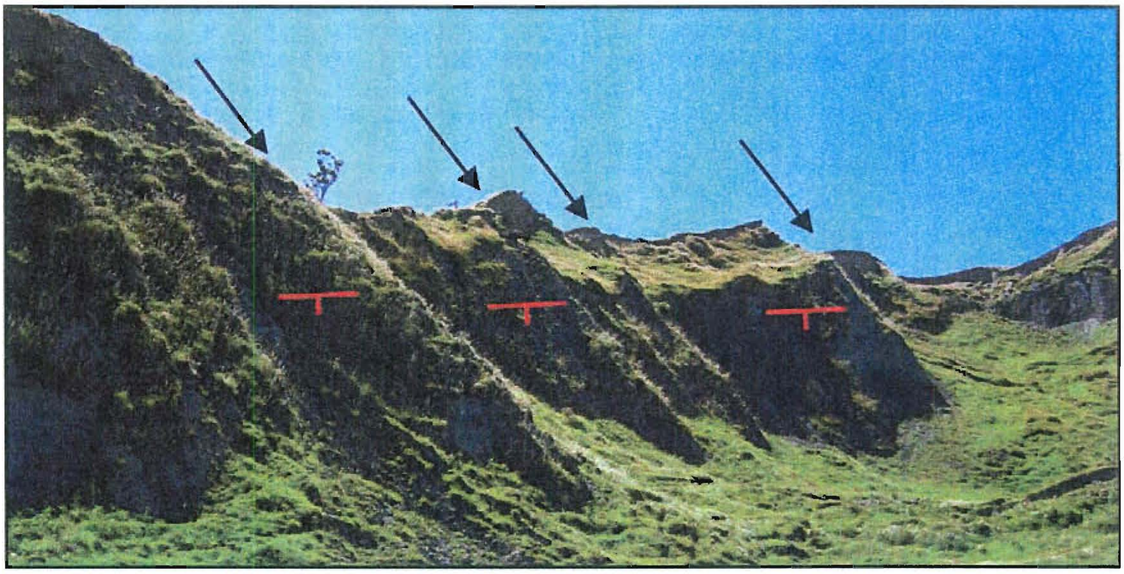

Figure 4.12: View of the right lateral scarp of the Amphitheatre Landslide, illustrating how defect control on landslide block release affects scarp geometry. The regular "sawtooth" scarp profile (sloping down left to right) directly reflects the orientation of a dominant joint set. Photograph taken looking north east from 6138150N 2843650E (NZMG 260 series map sheet V22).

Hydrogeological conditions

Little information is available on the hydrogeological conditions in the study area catchments. Of prime importance to a slope stability analysis is both the level of the free ground water table and the presence of confined aquifers. As the streams in the study catchments are deeply incised, and ridge tops are characteristically narrow, it is unlikely that the free water table would be near the ridge top and hence is unlikely to affect the failure of the Amphitheatre Landslide in the current setting. The hydrogeological regime at the time of landslide initiation is likely to have been different and the free water table may have had more influence on the stability of the slope when it was near the stream level. The occurrence of occasional seeps and springs near ridge tops indicated that there is water flow in these areas, however, no specific confined aquifers have been delineated and ridge top seeps are inferred to relate to increased permeability in fault zones.

A significant aspect of the hydrogeology in respect to slope stability in this study area relates to the dilation of the rock mass mentioned in the previous section. Rock mass dilation allows the development of secondary permeability or a network of fluid flow paths (as oppose to the primary permeability of the intact rock material) and this can allow rapid infiltration of water to the basal failure surfaces of partly displaced blocks. This secondary infiltration is only related to a rock mass that has been disturbed and dilated, so does not affect the intact rock mass that is of interest in terms of initial landslide triggering, but rather affects the behaviour of a slope following an initial displacement (e.g. Ponui and Waipoapoa Landslides, Pettinga, 1987a, 1987b).

4.2.3 Model of landslide development

The evolution of the Amphitheatre Landslide can be considered in terms of pre-existing factors within the rock mass which define the geometry and failure mode of specific landslides, and tectonic and climatic forcing factors which define the occurrence of slope failure on a catchment-wide scale.

The role of tuffaceous horizons in defining the basal failure surface of deep-seated landslides in the Makara Formation has been discussed. The Amphitheatre Landslide is failing on two levels defined by such critical stratigraphic horizons, however, the slope failure will not occur until the potential failure surface is exposed and overburden rock is laterally released. The incision of stream networks, driven by long-term tectonic and climatic forcing, exposes the stratigraphic succession and the thin, weak, sheared tuffaceous layers contained within it. The Amphitheatre Landslide would have initiated once the critical stratigraphic horizons defining the two basal failure surfaces of the landslide complex had been successively exposed at stream level.

The mode of slope failure that has allowed the Amphitheatre Landslide complex to develop involves moderate size blocks of intact material, the geometry of which are defined by the basal bedding parallel shear surface and the lateral release surfaces coincident with low or no strength joint sets. The size of these blocks is hence controlled by the stratigraphic spacing of critical stratigraphic horizons and the spacing of joint sets. The partially detached block in the lower level of the Amphitheatre Landslide represents a good example of a failure of this type (Figure 4.8 and Figure 4.13) and from this a block size can be defined based on both the plan geometry (~40x50 m) indicative of typical defect spacing where release is likely to occur, and the critical stratigraphic horizon spacing (100 m). The detached block has a slightly irregular plan geometry, and based on surveying of this plan geometry and projection of the basal geometry, the detached block has a calculated volume of approximately 3.2 x 10^5 m^3.

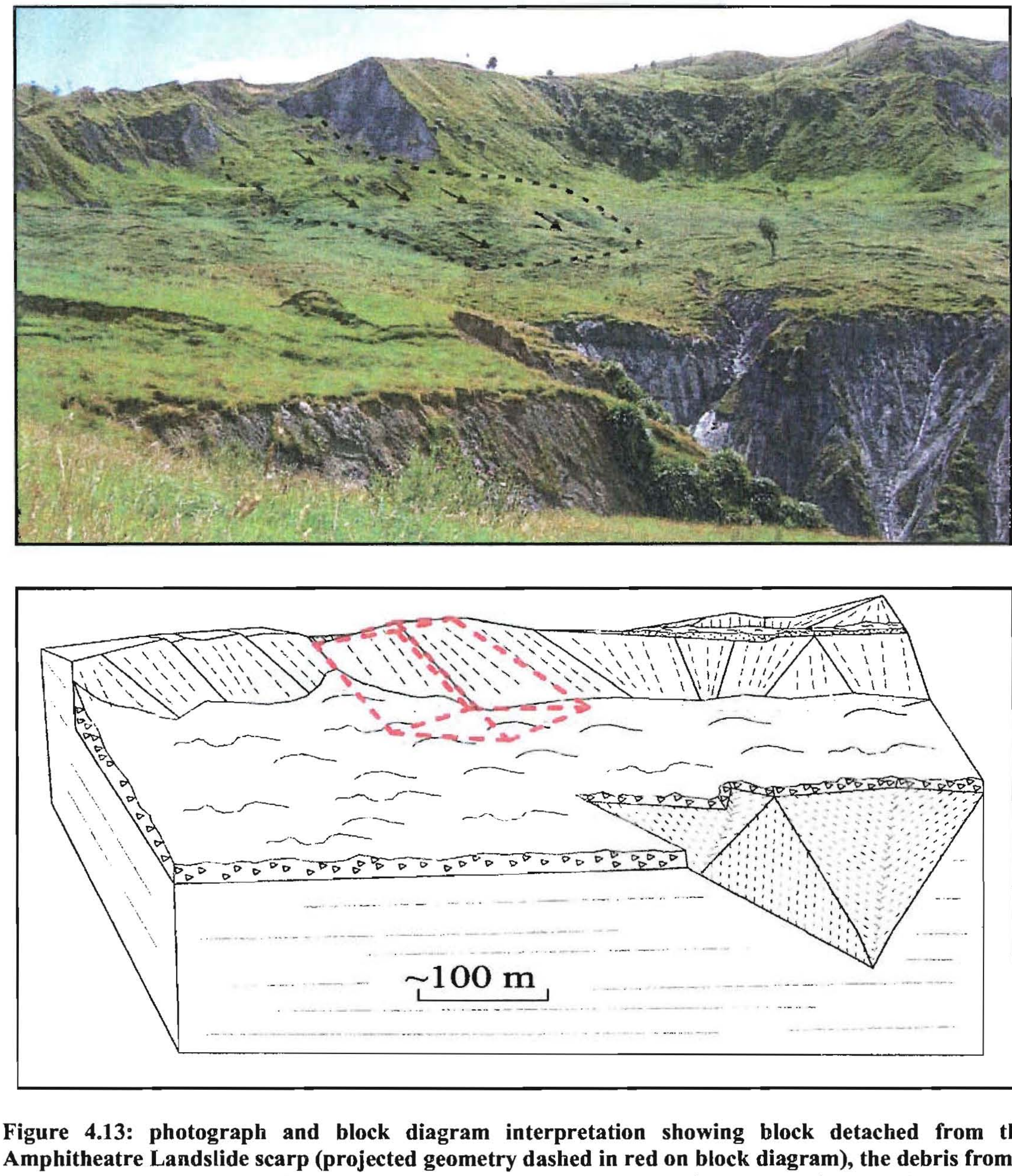

Figure 4.13: photograph and block diagram interpretation showing block detached from the Amphitheatre Landslide scarp (projected geometry dashed in red on block diagram), the debris from a collapsed block failure (outlined on photo) and the debris coverage of the basal failure surface (defined as a continuous layer on the block diagram). Photograph taken looking north east from 6138000N 2843450E (NZMG 260 series map sheet V22).

Both the initial development of the landslide at stream base level, and the subsequent enlargement of the landslide complex as it is progressively perched higher up in the catchment due to ongoing base level lowering, is defined by the detachment of blocks of this nature. Adjacent to the detached block in the lower level of the Amphitheatre Landslide debris indicates the rapid degradation and subsequent collapse of a similar sized block (Figure 4.13), and this is confirmed by vertical air photo interpretation. The degradation and full failure of blocks such as this directly relates to the rock mass dilation at the initial stage of failure (possibly caused by earthquake ground motion) and the subsequent secondary network of fluid infiltration paths which allows water access to critical zones in the slide mass and accelerates slaking and material degradation. Once these detached blocks have fully degraded they are entrained with material being transported across the basal failure surface by the creeping debris flow and introduced as sediment load into the fluvial network.

The development of the Amphitheatre Landslide complex over time can be described by the following sequence of events:

1. Incision of the Ponui Stream exposes a critical stratigraphic horizon which defines the upper level failure surface.

2. An initial landslide block defined by the depth of the critical stratigraphic horizon and spacing of conjugate joint sets is displaced, possibly due to earthquake ground motion dilating the rock mass. The block subsequently collapses into degraded debris due to enhanced water infiltration in dilated defects and rock material degradation.

3. Gradually more blocks fail in sequence, exhuming a larger area over the critical stratigraphic horizon and define an arcuate head scarp. Degraded material from collapsed blocks is transported across the basal failure surface under the influence of gravity but as the surface is still at stream base level material cannot be removed as fast as it is produced and the majority of debris accumulates on the landslide surface.

4. A period of rapid base level lowering exposes a lower critical stratigraphic horizon which defines the failure surface of the lower level of the landslide complex. Block detachment over this surface allows excavation of the lower landslide level which

gradually removes part of the higher upper level. Now that the upper level is above stream base level, debris can be transported onto the lower level and the upper level starts to obtain a bench like morphology.

5. A further period of rapid base level lowering leaves the lower level of the landslide complex perched above stream base level and as debris is now able to be transported away by the fluvial system at a similar rate to production the lower level obtains a bench like morphology.

6. The upper level continues to retrogress until it approaches the ridge crest. At this point activity decreases and the upper level attains a degraded form and pseudo-stable state. The lower level continues to excavate and progressively remove the upper level.

In the current configuration of the Amphitheatre Landslide complex, the lower level remains active and the upper level is pseudo-stable. The critical stratigraphic horizon recognised stratigraphically below the landslide complex may be inferred to define an incipient level of the Amphitheatre landslide complex. Some minor failure has already occurred on this critical stratigraphic horizon, and it is inferred that with a further period of accelerated stream incision the surface may become excavated to the point where it becomes a third and lower level of the Amphitheatre Landslide complex.

4.3 Chapter summary

The two landslides chosen from the selected field sites in North Canterbury and Hawke's Bay are representative of a broader population of landslide failure types commonly seen in New Zealand soft rock terrain. Specifically these landslides are both failing on thin, pre-sheared, stratigraphically and lithologically controlled surfaces which are inferred to be very common, if not ubiquitous, with deep-seated translational landslides in soft rock terrain.

Definition of specific failure controls on slope failure, such as joint sets, stratigraphically controlled failure surface strength, and the mode of initial failure of the landslide defines a model which will allow a quantitative assessment of slope stability. This slope stability assessment will allow consideration of the possible triggers for these specific landslides by assessing sensitivity to external factors. The triggering mechanism of these landslides can be inferred to apply to the wider population of landslides that the selected failures are considered to represent.

Chapter Five

5.0 Slope stability modelling

Computer based modelling of slope stability is particularly useful as it allows assessment of the sensitivity of the stability of a slope by variation of specific parameters. With well constrained and well defined slope parameters such as geometry, critical material strength and/or rock mass properties it is possible to consider sensitivity to key influences such as varying hydrologic conditions and seismic ground motion. Duncan (1996) provides a succinct overview of slope stability modelling methods. Three approaches are used in this study to consider the influence of earthquake generated strong ground motion on the stability of a slope.

- Static stability modelling refers to the stability of the slope under aseismic conditions. If a slope fails under static conditions, hydrological or anthropogenic factors are likely to play a role in initiating slope destabilisation.

- Pseudostatic stability modelling, which considers the effect of an earthquake on a slope as horizontal and/or vertical acceleration. Ground acceleration is represented as a static body force acting on a static slope stability model and subsequently this method only considers the influence of the strong ground motion at an instant in time.

- Dynamic stability modelling, which refers to the consideration of how a slope responds for the duration of an earthquake acceleration time history.

A combination of static, pseudostatic and dynamic stability modelling is useful to describe the sensitivity of a slope and its behaviour during an earthquake event. The main purpose of this is to assess the stability of the two representative landslides selected for this study, the Ella Landslide in North Canterbury and the Amphitheatre Landslide in Southern Hawke's Bay, using computer based slope stability modelling methods. This chapter addresses a primary objective of this study, to quantitatively assess the role of strong ground motion as a triggering mechanism for large prehistoric soft rock landslides. Following background information on the slope stability modelling methods employed the stability of the Ella Landslide and Amphitheatre Landslide will be considered in turn.

5.1 Slope stability modelling methods

5.1.1 Static stability analysis

The most applicable method for modelling the stability of the Amphitheatre and Ella slope failures is a planar failure model applicable to rock slope stability assessment, principally based on the numerical methods of Hoek and Bray (1981). The slope stability modelling program RocPlane© developed by Rocscience Inc. allows for the two-dimensional assessment of the stability of a planar wedge or block slide. RocPlane© uses the limit equilibrium method (Rocscience Inc., 2001), in which the stability of a slope is given as the ratio of the total forces resisting down slope sliding to the total forces driving down slope sliding (defined as the factor of safety, Fs). The RocPlane© model assumes that a block is sliding on a failure plane that dips at less than the average slope angle (daylights within the face) and strikes approximately parallel to the slope face. The model considers a slice of unit width in the direction of failure, and assumes that the lateral release surfaces are insignificant with respect to the sliding resistance acting on the block/slice.

Figure 5.1 shows a simple planar wedge geometry, for which it is possible to vary the slide mass dimensional values in order to define the geometry applicable to specific situations. Other variable parameters include the strength properties of the failure surface (in terms of the Mohr Coulomb failure criterion), ground water pressure, and seismic load.

RocPlane© allows for either deterministic or probabilistic analysis. For deterministic analysis, parameters are given a single mean value and a single factor of safety value (Fs) is derived. In a probabilistic analysis, values are assigned a statistical distribution (usually normal), defined by a mean value, standard deviation, and a relative minimum and maximum. For probabilistic analysis the output is a histogram (plotting iterations of block stability) and a probability of failure is defined as the ratio of the area where Fs < 1 to the area where Fs > 1 (Hoek, 1998). As some parameters are inherently uncertain (e.g. strength data and hydrological conditions associated with any prehistoric slope failure), a probabilistic slope stability analysis is a useful supplement to a deterministic analysis.

The most likely variable that would cause a deep-seated landslide to occur in the natural environment is changing hydrological conditions. As the primary aim of stability modelling in this study is to assess the likelihood of an earthquake trigger, it is crucial to have a realistic ground water regime to define a representative static slope stability model and ensure that the possibility that the selected slopes may have been triggered by elevated pore pressures is comprehensively considered.

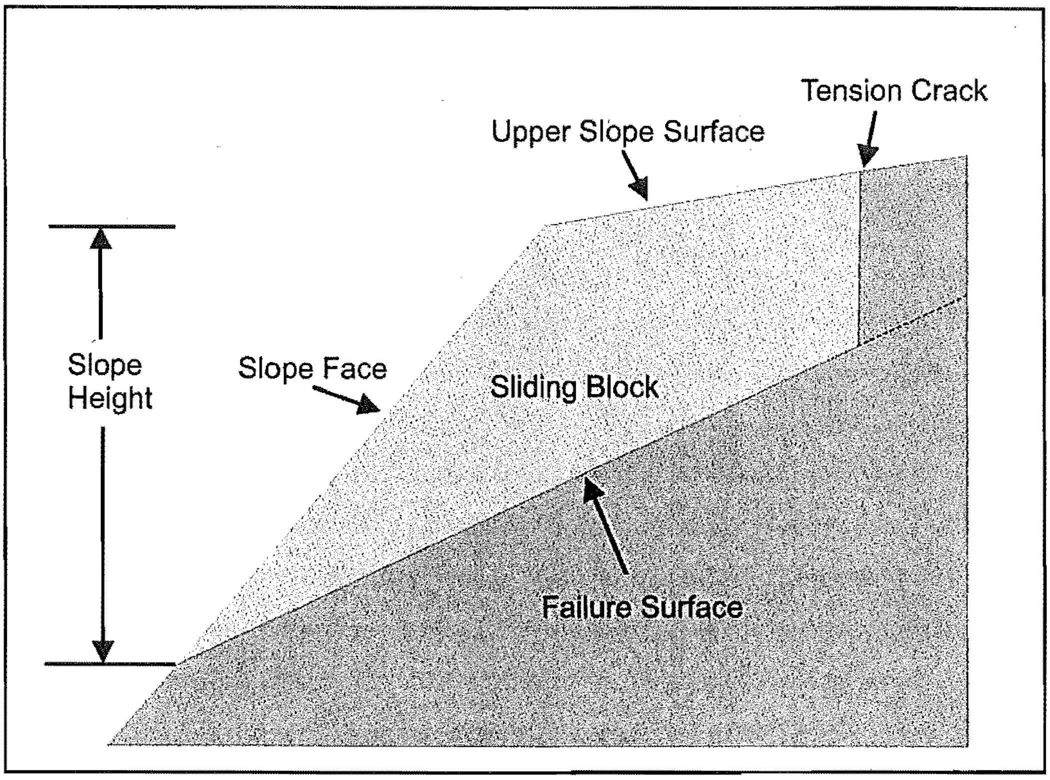

Figure 5.1: Basic geometry of a planar wedge failure as defined by the slope stability modelling package RocPlane©. Failure of this planar wedge would occur from right to left.

Hydrogeological behaviour of jointed rock masses

Groundwater commonly has a role in the destabilisation of rock slopes and in jointed rock masses and the primary causes of this are: i) pressure variation acting on discontinuities; and, ii) alteration and/or transportation of materials within the rock mass such as defect infill, (Giani, 1992). There are four models of pore pressure distribution within a slope that are considered by RocPlane©, focused at the toe or the middle of the failure surface (Appendix II).

The permeability of alternating sandstone and siltstone soft rock sequences has been reported to be in the range of 10^{-4} – 10^{-8} ms^{-}1 (Brown, 1974). Measurement of the permeability of Tokama Siltstone in North Canterbury agrees with this ($\leq 1.28 \times 10^{-8}$ ms^{-1} by Geotech Consulting Ltd., 2002), and the permeability of the Makara Formation in Southern Hawke's Bay is also inferred to fall at the lower end of this range. These permeability values relate to the properties of the intact rock material (the primary permeability) and do not reflect the higher permeability of discrete clean sand beds within the succession, or rock mass defect permeability (secondary permeability). The occurrence of higher permeability sand beds within low permeability units has a specific bearing on deep-seated bedrock slope failures, as elevated pore pressures can develop within such horizons and cause them to act as failure surfaces (e.g. Ker, 1970; Pettinga and Bell, 1992). The hydrological regime of the two study sites was discussed in Chapter 4, and the influence of both primary and secondary permeability on slope failure within the two field sites will be discussed in subsequent sections of this chapter.

5.1.2 Pseudostatic stability analysis

A pseudostatic analysis considers the behaviour of a slope during an earthquake by representing earthquake shaking as inertial forces acting through the centroid of the failure mass (Kramer, 1996). The method is useful to consider the binary response of a slope to the peak ground acceleration during an earthquake, i.e. did the slope fail or not? Ground acceleration is generally only applied in the horizontal direction and the vertical component neglected, as the (upward) vertical pseudostatic force reduces both the driving force and the resisting force. If, for a statically stable slope, the magnitude of ground acceleration is increased until the slope reaches unity (factor of safety = 1.0) then that level of ground acceleration can be considered the minimum required to induce failure (termed the critical or yield acceleration). This is not necessarily equivalent to the peak ground acceleration for that earthquake, but an earthquake which triggers slope failure must achieve at least that level of ground motion.

A pseudostatic analysis is relatively straightforward to conduct, especially when a static analysis of a slope has been previously undertaken. The approach has limitations, however, as it considers the complex and dynamic inertial forces which occur over a specific duration during an earthquake as momentary pseudostatic inertial forces (Kramer, 1996). There are several (dynamic) methods of seismic slope stability available which consider the permanent deformation of a slope as result of an earthquake event and the method that is in most widespread use, the Newmark sliding block analysis, is considered here.

5.1.3 Dynamic stability analysis

Newmark (1965) developed an analysis method for assessing the seismic stability of embankment dams based on a simple model of a block sliding on an inclined plane (hence referred to as the "Newmark analysis"). The method has subsequently been modified for application to natural slopes (Wilson and Keefer, 1983) and is used extensively to assess dynamic slope behaviour (e.g. Jibson and Keefer, 1993; Jibson et al., 2000; Romeo, 2000; Shou and Wang, 2003; Murphy and Mankelow, 2004). Jibson (1993) provides a comprehensive review of the method and its applicability to natural slopes.

The Newmark analysis treats a landslide as a rigid-plastic body and is hence applicable to translational block slides which will experience no internal deformation. The displacement of such a block (termed the Newmark Displacement) can be quantified by double-integrating all parts of a strong motion record which fall above the yield acceleration (defined in the previous section) for a particular slope (Figure 5.2).

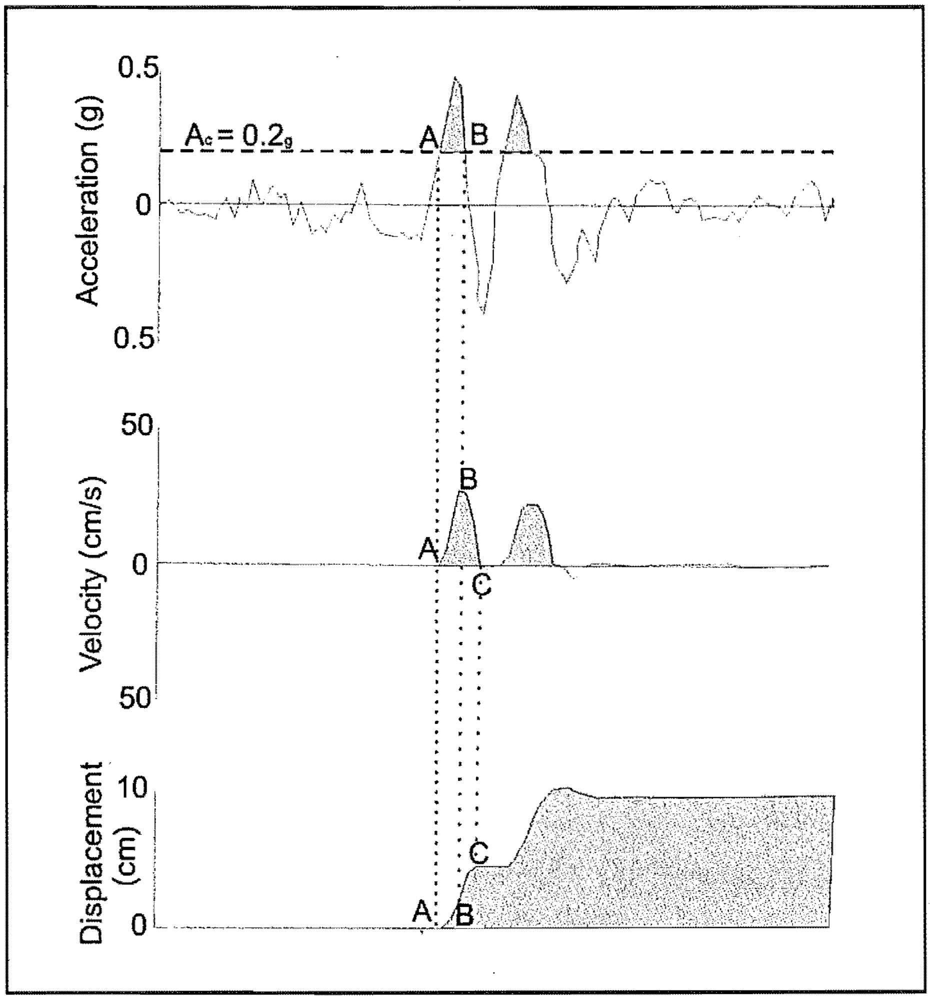

Figure 5.2: Calculating Newmark Displacement from a strong motion record for a given critical or yield acceleration (Ac). Modified from Wilson and Keefer (1983).

This requires that a digitized strong motion record is available which represents the ground motion of a particular earthquake at that particular site. The Newmark analysis program of

Jibson and Jibson (2003) includes 2160 strong motion records from 29 earthquakes, and the program also allows the user to input digitised strong motion records for a particular region (in New Zealand these are available in digitised form from as far back as 1966). To obtain a true assessment of the response at a particular site to a specific earthquake requires an actual strong motion record for that event, recorded at that site. For pre-historic slope failures this is obviously impossible and so for a given scenario (inferred earthquake a given distance from a specific site) a strong motion record must be obtained that represents the magnitude of the inferred earthquake and was recorded at an equivalent distance from the epicentral location by a strong motion recorder founded on equivalent geological materials. As the location of such a specific strong motion record may be problematic from brief historical databases, a simplified empirical version of the Newmark analysis method has been developed based on available strong motion records (Jibson and Keefer, 1993; Jibson et al., 2000).

The simplified Newmark method recognises the limitation of the peak ground acceleration as a descriptor of strong ground motion and considers the relationship between slope displacement and the Arias Intensity (a velocity defined by the integration of the entire duration of earthquake ground motion). The method requires the user to input the critical acceleration and the Arias Intensity to allow calculation of the Newmark Displacement. If a critical acceleration is known, and a Newmark Displacement is assumed, then this algorithm could be iteratively used to predict the Arias Intensity required to induce failure. The estimate is made using the following regression equation:

$$\log D_n = 1.521 \log I_a - 1.993 \log a_c - 1.546 \qquad (1)$$

where D_n is Newmark Displacement in centimetres, I_a is Arias Intensity in metres per second, and a_c is the critical acceleration given as a fraction of the acceleration of gravity (g) in m/s. This equation was developed by conducting rigorous Newmark integrations on 555 single-component strong-motion records from 13 earthquakes for several discrete values of critical acceleration. The regression model has an R^2 value of 83% and a standard deviation of 0.375.

The predicted Arias Intensity can then in turn be used to predict the required earthquake magnitude. Wilson and Keefer (1983, cited in Jibson, 1996) developed a relationship between Arias Intensity, earthquake magnitude, and source distance:

$$\log I_a = \mathrm{M} - 2\log R - 4.1 \qquad (2)$$

where I_a is Arias Intensity in metres per second, M is moment magnitude, and R is earthquake source distance in kilometres.

These relationships are developed from international case studies and each case reflects the response of a specific geological, tectonic and geomorphic situation. It is likely that caution needs to be used when applying these relationships, as the response of a given site may not necessarily be represented in the data set used to derive the relationships.

5.2 Ella Landslide stability modelling

The geological model for the failure mode and geometry of the Ella Landslide is detailed in Chapter 4. Here the analysis of the stability of the landslide with respect to the likely triggering mechanism(s) is considered using a combination of static, pseudostatic and dynamic slope stability modelling techniques.

5.2.1 Static Stability of Ella Landslide

The static stability of Ella Landslide can be assessed using the computer based slope stability modelling package RocPlane©. To define the stability of the pre-failure slope a deterministic analysis is undertaken based on the model of the pre-failure slope geometry shown in Figure 5.3.

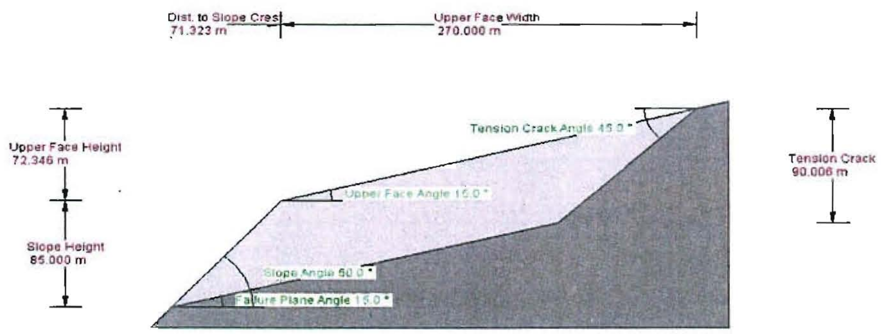

Figure 5.3: Slope stability model for the pre-Ella Landslide slope, based on the slope configuration presented in Figure 4.5.

Slope sensitivity to parameter variation

The sensitivity of the model to variation in specific parameters indicates which of these parameters have a significant impact on overall slope stability. With the slope geometry shown in Figure 5.3, the mean strength values listed in Table 5.1, and considering a drained slope the sensitivity of the model to slope height, slope angle, upper face width, failure plane angle and material friction angle can be assessed (plots shown in Appendix III). The static factor of safety for the slope with mean values for these parameters is 1.3, and in terms of slope sensitivity a variation in the factor of safety of at least 0.1 over the given range of values is considered to be significant.

The parameters which significantly affect slope stability are the failure plane angle and the friction angle. The dip of the Tokama Siltstone defines the failure plane angle and this is considered to be well constrained by field mapping. Without considering hydrological or seismic influences (these will be considered later in this chapter) variation in laboratory derived shear strength (friction angle) values for the Ella Landslide failure surface material is considered to be the parameter that has the most potential to influence the stability model and this is supported by published literature (e.g. Bromhead et al., 2002).

Parameter	Mean Value	Standard deviation	Relative minimum/maximum
Friction angle	18.5 degrees	1	±3 degree
Cohesion	0.25 t/m^2	0.1	±0.25 t/m^2
Unit Weight	2.0 t/m^2	0.1	±0.2 t/m^2
Failure plane angle	15 degrees	0.5	±2 degrees
Slope height	85 m	17	±20 m
Upper slope width	270 m	35	±50 m

Table 5.1: Values for parameters used in the stability analysis for Ella landslide. For probabilistic analysis purposes all parameters have a normal distribution.

To incorporate the three dimensional nature of the landslide in the analysis, the stability can be considered for several sections through the slope. All sections considered are shown in Figure 5.4, spaced at 100 m centres. While there is some variation in stability between these sections it is not considered to be significant. The parameters which vary are the slope height and upper face width and the model has been shown to be relatively insensitive to these. Variation of these two parameters between the five sections affects the factor of safety for the slope by less than 0.001. The central section (Section 3) is hence considered to be representative of the stability of the pre-Ella Landslide slope. An assumption of the two dimensional RocPlane© model is that edge effects are negligible. The use of three dimensional analysis is commonly considered unnecessary in engineering practice as two dimensional analysis is considered conservative (Duncan, 1996), however, the reality of three dimensional slope failures is that variation of parameters throughout the landslide footprint are likely (Bromhead et al., 2002). The style of rock mass defect controlled slope failure considered in this study is assumed to have lateral release surfaces defined by strength free joint sets and this is clearly supported by field evidence in the Hawke's Bay field area (well defined joint sets and slopes defined by joint set orientation), and is supported by the linearity of the lateral scarps of the Ella Landslide. Given this assumption, and the ability to vary parameters (e.g. material strength) within a two dimensional analysis, a fully comprehensive three dimensional slope deformation model is not considered to be necessary or practical given time constraints for this study.

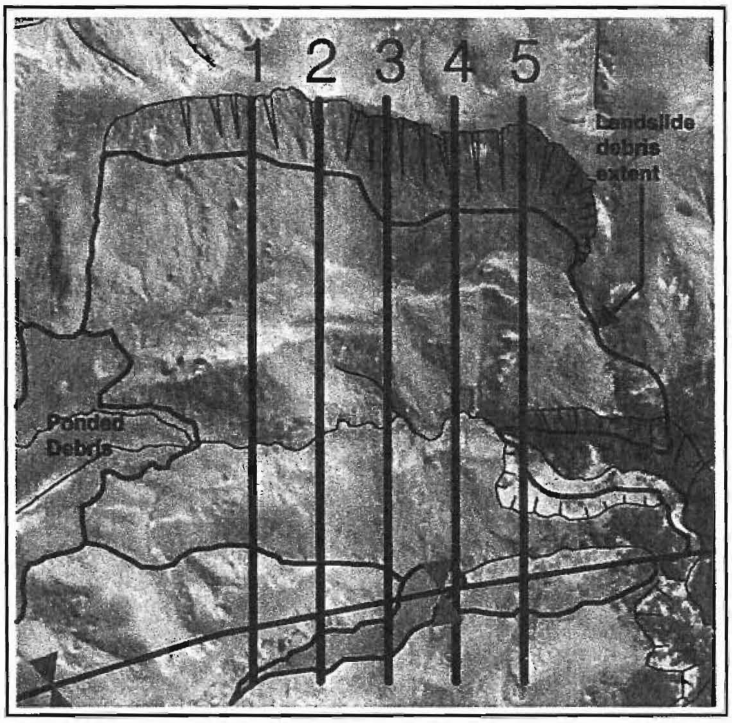

Figure 5.4: Distribution of slope sections considered for stability analysis of the Ella Landslide. Section three is used for the main analysis and the slope profile is shown in Figure 5.3.

To allow consideration of the variability of all parameters within a single slope stability analysis, a probabilistic analysis is undertaken. Using the parameter distribution presented in Table 5.1, with a dry slope, the probability of failure of the Ella Landslide is 0% (Figure 5.5).

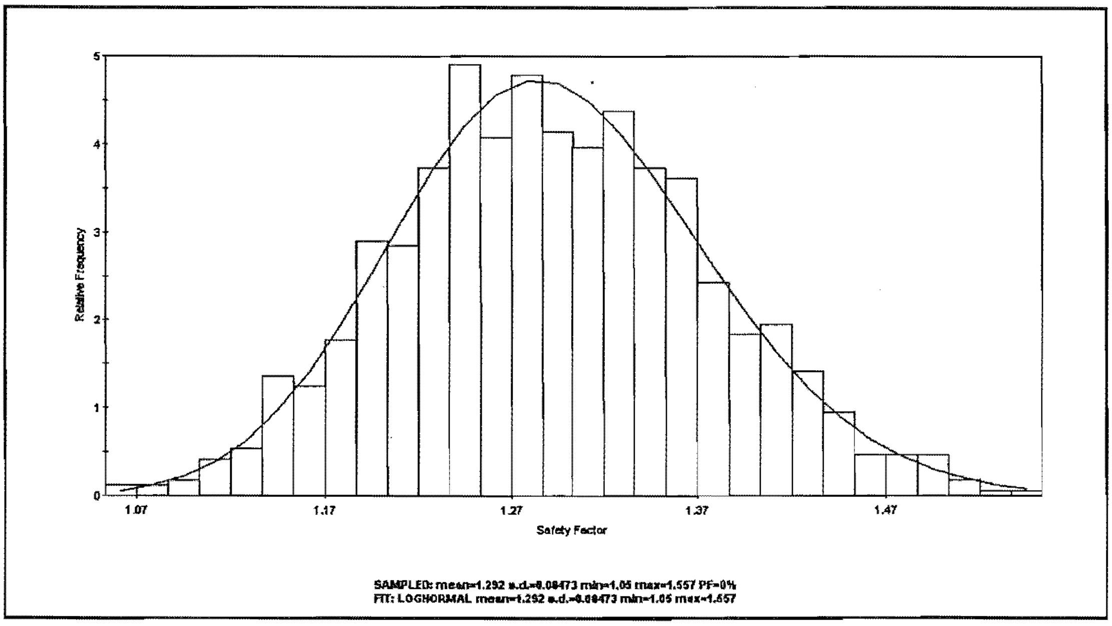

Figure 5.5: Factor of safety distribution showing a probability of failure of 0% for a Probabilistic analysis of the Ella Landslide.

Both the deterministic and probabilistic methods of stability analysis used indicate that the slope is likely to be stable in static, drained conditions.

Hydrological influence on the failure of Ella landslide

Hydrogeological conditions in Kate Valley are discussed in Section 4.1.4, and it is considered unlikely that the unconfined groundwater table rises much above local stream level in the cross valley profile. As the toe and failure surface of Ella Landslide are considered to have been primarily above stream level at the time of failure, the slope is unlikely to be significantly affected by groundwater induced pore-pressure. The sensitivity of the slope to varying degrees of saturation can easily be assessed, however, and RocPlane© allows four different water pressure distribution models (Appendix II). In a worst case hydrological scenario the profile of the water table could rise towards the ridge top from stream level, and Figure 5.6 illustrates this situation showing the water table (curved, dashed line) and associated pressure distribution on the failure surface.

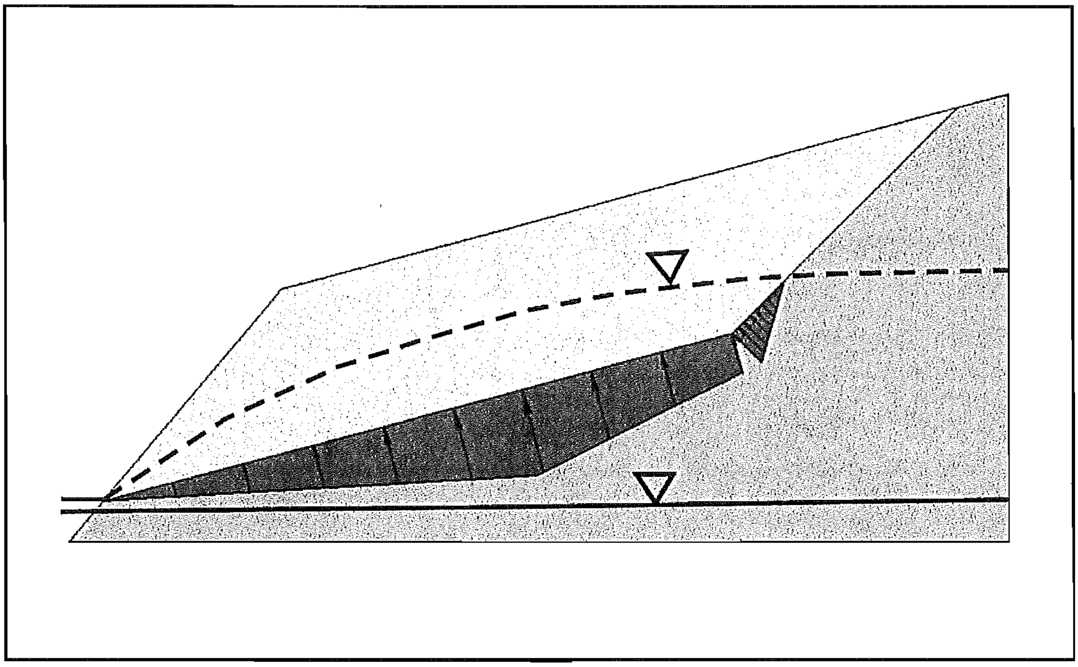

Figure 5.6: Possible ground water level for the Ella Landslide slope at the time of failure which could cause the slope to be 56% saturated (dashed line) and the preferred groundwater configuration (solid line). The dark grey area and arrows show the pore pressure distribution on the failure surface related to the dashed line groundwater regime.

If the sensitivity of the slope to variation in groundwater (water pressure concentrated at mid-height) is considered, the slope is at unity (Fs = 1.0) when it is 56 % saturated (Figure 5.7) and in a probabilistic analysis this equates to a probability of failure of 48 %. The ground water regime which would allow the slope to be 56% saturated is shown in Figure 5.6.

Based on the known ground water trends in Kate valley, it is considered unlikely that ground water pressure would build up to 56 % in this slope. The only potential for such high pore pressure build up is considered to be on the clean sand beds which define perched aquifers. For this to occur in Ella Landslide would require the presence of a sand bed (such as the one indicated in Figure 4.4) stratigraphically adjacent to the critical stratigraphic horizon which is inferred to have acted as the basal failure surface for the landslide. No such sand bed occurs in the exposure from which this critical stratigraphic horizon has been characterised (Figure 3.2), however, the existence of such a bed cannot be completely discounted due to the lack of subsurface data from within the landslide footprint. Field evidence suggests that there was no perched water table involved in the Ella Landslide Failure, however, and this mechanism of pore pressure build up will not be considered further.

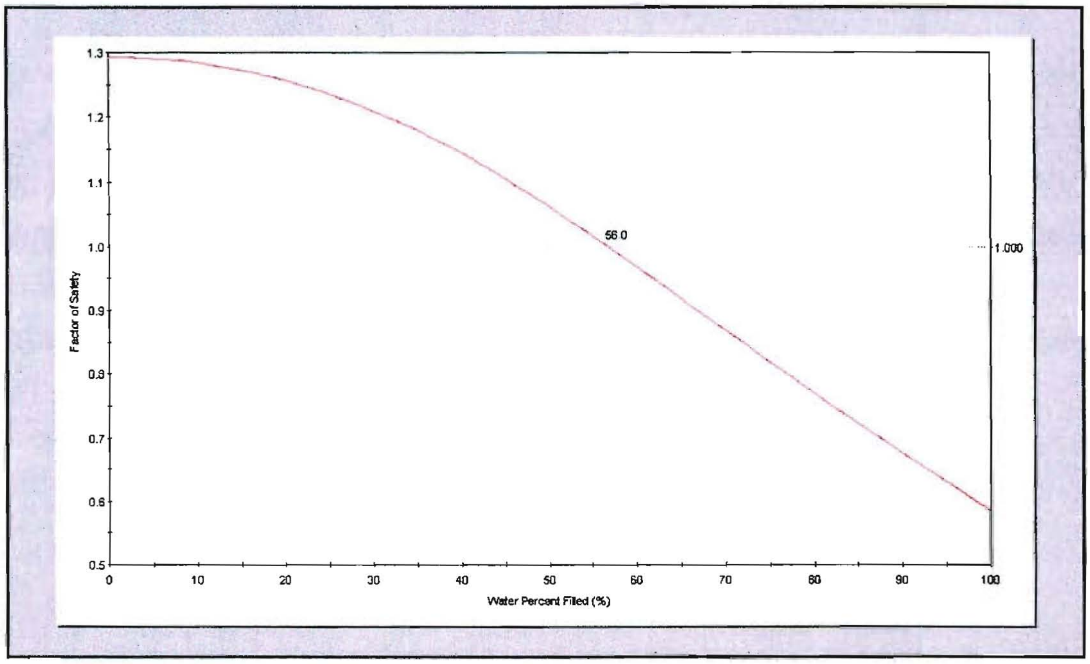

Figure 5.7: Sensitivity of Ella Landslide to hydrological conditions in a probabilistic analysis, using parameters defined in Figure 5.3 and Table 5.1.

Failure under purely adverse hydrological conditions is considered to be unlikely, and as the actual groundwater conditions at the time of failure are unknown, a realistic condition must be inferred. With the toe (and failure surface) of Ella Landslide inferred to be above stream level at the time of failure, the groundwater table in this area having a predominantly flat profile and there being no evidence for the occurrence of perched water tables acting on the failure surface of the landslide, it may be assumed that the slope was fully drained at the time of failure.

Release surface strength

The computer package used for slope stability analysis (RocPlane©) considers that the slide mass releases on a tension crack. For the Ella slope this is used in the context of a jointed rock mass to model the release of the Ella Landslide mass on a defect surface with no strength. In the Tokama Siltstone there is limited field evidence to support the occurrence of laterally extensive defect surfaces (refer Chapter 3), however, in many documented cases of deep-seated bedding parallel failures in soft terrain in New Zealand, the head-ward landslide release mechanism is defined by rock mass jointing (e.g. Thompson, 1981; Pettinga, 1987a, 1987b; Pettinga and Bell, 1992).

While it is considered to be most likely that the Ella Landslide released on a low or no strength surface, if total or partial rupture through intact Tokama Siltstone was necessary for slope failure to occur then this might significantly affect the stability of the slope and the effect of this should be considered. One way to approach this is to use a limit equilibrium program such as Slide© (also published by Rocscience Inc.) which allows the user to define a layered slope geometry and define a specific failure surface. The slope model is defined with the geometry shown in Figure 5.8, using material and geometric parameters defined in Figure 5.3 and Table 5.1, and using strength values for the intact Tokama Siltstone as cohesion of 176 kN/m^2 and a friction angle of 37° (as defined in Table 3.1).

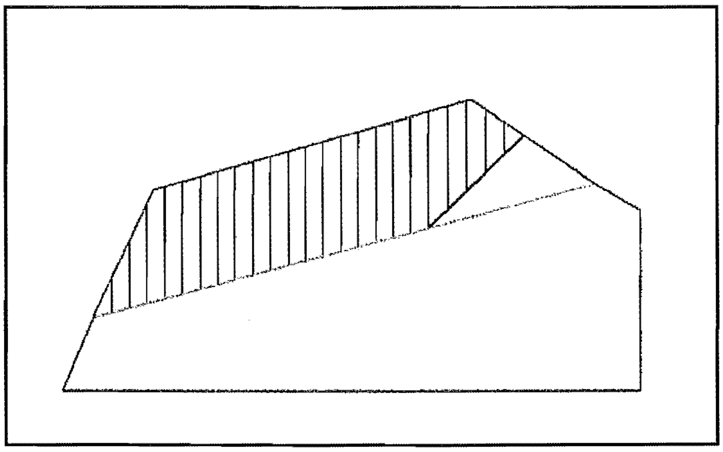

Figure 5.8: Basic slope geometry for the Ella Landslide as defined in the stability package Slide©. The vertical slices define the extent of the modelled slide block which has its basal failure surface defined by the critical stratigraphic horizon (light grey line) and head scarp release through intact Tokama Siltstone.

The failure surface defines the base of the block as sliding on the low strength clay layer, while the release surface (which defines the present day head scarp) ruptures through intact material. Surprisingly, this model indicates similar slope stability to the model of the slide block releasing on a tension crack surface, and results in a factor of safety of 1.3 (equivalent to the static Fs defined by RocPlane©). With the purely translational geometry of this landslide it might be considered that the release on the headward surface would be solely dependant on the tensile strength in the dip direction (e.g. Maslov et al., 1981) and hence the shear strength of the intact material (which limit equilibrium model such as Slide© consider) would not affect the strength of this part of the slope in a purely translational slide. As no testing of the Tokama Siltstone tensile strength has been undertaken, the strength might be inferred from published data of other weak rock lithologies as being in the order of 0.5 – 1 MPa (Waltham, 1994). By considering the thickness of the slide block perpendicular to bedding the effect of the tensile strength of the rock material on slide block release can be estimated. As RocPlane© uses strength units of tonnes/square metre, the tensile strength of the material might be considered to be in the range of 50 – 100 t/m^2. The block thickness is approximately 82 m and this equates to an overall resistance to sliding (in terms of a point load) for a unit slope slice of 4100 – 8200 t/m. To consider the effect of this strength on the stability of the slope it is possible to apply a point load against the direction of sliding. It is inferred that the slide block release might be only partly resisted by intact material strength and partly controlled by some form of defect (Chapter 4). In this case it is thought to be prudent to consider the stability of the slope with the lower value of material strength to account for this. The resulting factor of safety for this analysis is 2.4.

The stability of any prehistoric slope failure is inherently uncertain. A representation of the slope geometry, rock mass and material strength may be fairly well defined but uncertainties will remain with these parameters and specifically with hydrological conditions. For the Ella Landslide, the existence of the failure surface above stream level, and hence the water table, means that the slope is modelled as dry. In this condition the slope is stable and would require some external perturbation to trigger failure (e.g. a seismic event). For this statically stable situation it is possible to consider the effect of earthquake ground motion on the slope using a pseudostatic analysis.

5.2.2 Pseudostatic stability of Ella Landslide

Pseudostatic analysis considers the influence of an earthquake in terms of constant ground acceleration. It is possible to look at varying levels of ground acceleration with respect to the slope model discussed for the Ella Landslide and assess what level of ground motion would be required to reduce the factor of safety (Fs) to 1.0 (unity). As discussed we have two scenarios for the Ella Landslide slope, one considering a release on a defect with no strength whatsoever (Fs = 1.3) and another considering the tensile strength of the intact material (Fs = 2.4).

For Fs = 1.3 the minimum ground motion required to induce slope failure is 0.07g, while for Fs = 2.4 this rises to 0.19g. This means that, depending on what is the reality of the slope model this is the minimum amount of ground motion required to induce failure in the slope. It is possible to use published empirical relationships to infer what magnitude of earthquake would be required to produce this level of ground motion and compare this to potential earthquake ground motion in this area.

5.2.3 *Dynamic Stability of Ella Landslide*

To truly consider the dynamic behaviour of a slope it is necessary to consider the nature of earthquake induced ground motion not as an inertial force (as in pseudostatic analysis) but as a variable force over time. The Newmark displacement (discussed in Section 5.1.3) defines the amount of landslide displacement which will occur over a specific earthquake acceleration time history (Jibson and Keefer, 1993). A rigorous Newmark analysis requires acceleration time history data for the actual earthquake which triggered a specific slope failure, however, for a landslide which occurred more than 5000 years ago this is difficult even to infer (due to variables such as source distance, foundation conditions and the nature of a specific fault rupture). For Ella Landslide it is interesting to look at the probable magnitude of earthquake which would have triggered slope failure as this provides information on the magnitude of seismic events which might be considered critical to the development of the landscape. Two published empirical relationships (discussed at the beginning of this Chapter) which consider the relationship between Arias Intensity, yield acceleration and Newmark Displacement, and between Arias Intensity, earthquake magnitude and source distance might be used to infer the magnitude of earthquake required for slope failure using an inferred Newmark Displacement and the yield acceleration predicted from pseudostatic analysis. The inferred earthquake magnitude can be compared to maximum credible earthquakes from known earthquake sources in the region.

Earthquake sources for Kate Valley

The seismic hazard in Canterbury is well documented (Stirling et al., 1999, 2001) and the seismic potential for the Kate Valley is outlined in Table 5.2. This shows that even the upper predicted yield acceleration for Ella Landslide of 0.19g is certainly not unlikely and in fact this value may have been realized twice in historical time (Geotech Consulting Ltd, 2002).

Two possible earthquake scenarios for Kate Valley are identified by Geotech Consulting Ltd (2002). These are a Mw = 6.8 ± 0.3 rupture on the Omihi Fault at 4.2 km to the west-north west which could produce 0.65 – 0.69g and a Mw = 7.25 rupture of the Porters Pass-Amberley Fault Zone at a minimum 21 km distance producing ~0.36g at the Ella Landslide.

Return Period	50 yr	150 yr	475 yr	1000 yr
PGA	0.25g	0.37g	0.55g	0.7g
Continuous MM Intensity	8.0	8.5	9.0	9.3

Table 5.2: Strong ground motion potential in Kate Valley based on probabilistic seismic hazard modelling, from (Stirling et al., 1999).

Predicting the earthquake which triggered Ella Landslide

We can use the predicted yield acceleration and an estimated critical displacement to predict the Arias Intensity at the site during the earthquake which induced Ella Landslide to fail. We have calculated the yield acceleration as being either 0.07g or 0.19g, however, as the lower value is likely to be regularly exceeded, the upper vale will be considered here. There is no control on what the critical displacement of the Ella Landslide block might be, however, Jibson (1993) uses average critical displacements based on various case studies in the order of 5-10 cm for landslides in North America.

If the value of 10 cm is adopted for the Newmark displacement on the Ella Landslide, equation 1 (Section 5.1.3) predicts an Arias Intensity of 5.4 m/s for the 0.19g yield acceleration. Table 5.3 shows four fault rupture scenarios which could be considered to have produced large earthquakes in this area. Using equation 2 (Section 5.1.3), the magnitude calculated from the predicted Arias Intensity can be compared with the maximum credible earthquake for that fault.

The Hamilton and the Omihi Faults both have potential to produce an earthquake of sufficient magnitude to trigger the Ella Landslide to fail. The Hamilton fault was not included in the Pettinga et al. (1998) study as a potential earthquake source, but was considered by Geotech Consulting Ltd. (2002) in a conservative approach with respect to the seismic hazard at the site of the new regional landfill in Kate Valley. For this project the Hamilton Fault is considered to be an unlikely earthquake source. The North Canterbury Fold and Fault Belt extends up to 20 km offshore (Pettinga et al., 1998) and includes thrust faults oriented with the regional NE – SW structural grain. These faults are not thought to have been active during the Holocene and so do not need to be considered in this study. From comparison of the predicted moment magnitudes from this study with the maximum credible earthquakes for faults that are considered as potential earthquake sources (refer Table 5.3) the most likely earthquake source to trigger the slope failure is a rupture on the Omihi Fault. The Omihi Fault is considered to have ruptured in the last 10, 000 years (Stirling et al., 1999) and is inferred as the possible earthquake source for triggering the Ella Landslide.

Fault	Distance from site	Calculated Magnitude (this study)	Maximum Credible Earthquake for Fault
Hamilton Fault	2 km	Mw = 5.4	Mw = 6.2 (Geotech Consulting Ltd, 2002)
Omihi Fault	4.2 km	Mw = 6.1	Mw = 6.8 (Geotech Consulting Ltd, 2002)
Porters Pass – Amberley Fault Zone	21 km (minimum)	Mw = 7.5	Mw = 7.25 (Stirling et al., 1999)
Alpine Fault (Kaniere – Tophouse Segment)	99 km	Mw = 8.8	Mw = 7.7 (Stirling et al., 1999)

Table 5.3: Calculated earthquake magnitude (this study) vs maximum credible earthquake for four faults with respect to the Ella Landslide.

The empirical relationships used to predict the triggering earthquake for the Ella Landslide are developed from data for earthquakes and landslides in a wide variety of geological and tectonic settings and therefore may not be totally suitable for the specific situation in Kate Valley. Despite this uncertainty, stability analysis overall indicates that the slope was likely to be stable under static conditions and the level of earthquake shaking predicted agrees with the potential of local earthquake sources. If similar geomorphological and geological conditions to those occurring at the time of the Ella Landslide slope failure coincide with a rupture of the Omihi Fault then there is potential for a slope failure of similar magnitude to occur in the future.

5.3 Amphitheatre Landslide, Hawke's Bay

The development of the Amphitheatre Landslide as a retrogressive landslide complex rather than a single failure event (as discussed in Chapter 4), means that the stability of the landslide complex is defined by the stability of the individual block slides by which it has developed. Based on this model of landslide development it is considered that the displaced block on the head scarp of the landslide complex (Figure 4.8 and Figure 4.13) can be considered to be representative of the stability of the Amphitheatre Landslide complex as a whole. The material properties, geometry and mode of failure of this block are inferred to be representative of the numerous blocks which periodically detach from the retrogressing landslide scarp. Many landslides in the catchments adjacent to the southeastern margin of the Maraetotara Plateau are failing in a broadly similar retrogressive manner (this will be discussed further in the following chapter) and so the stability of the Amphitheatre Landslide is considered as representative of this style of landslide in these catchments.

5.3.1 Static stability

Using the slope stability modelling package RocPlane© the stability of the Amphitheatre Landslide block can be considered, based on the geometry shown in Figure 5.9 with the parameters listed in Table 5.4. As the geometry of the intact block is accurately defined, the parameters which are considered to have uncertainty are the material strength properties, hydrological conditions and the angle of the failure plane.

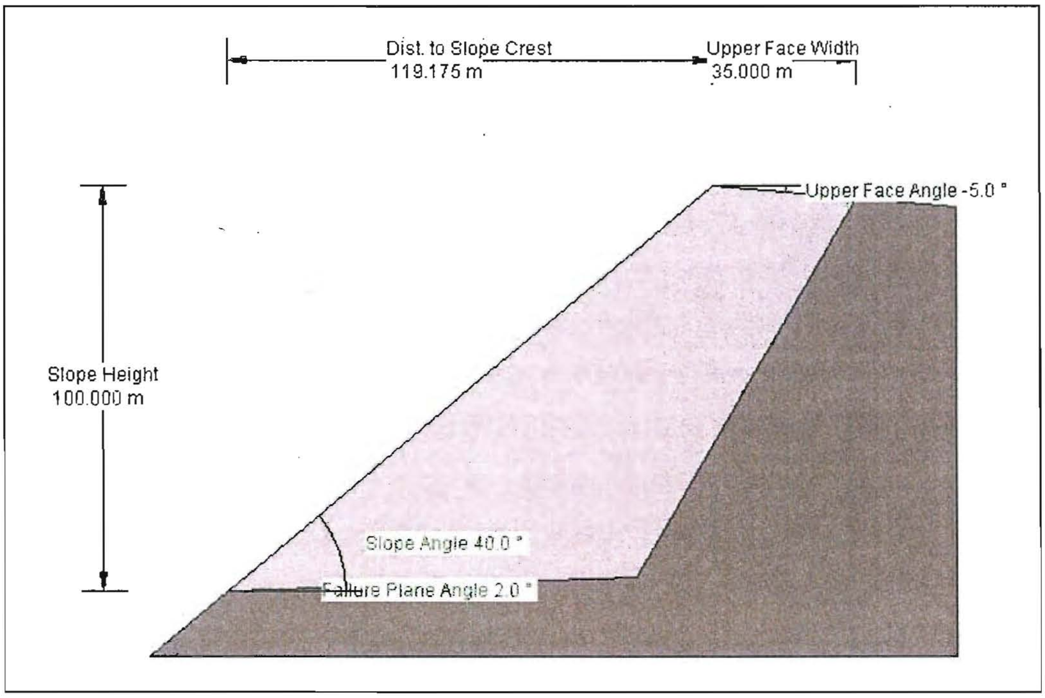

Figure 5.9: Slope stability model geometry for the Amphitheatre block. Mean failure plane angle is shown.

Parameter	Mean Value	Standard deviation	Relative minimum/maximum
Failure plane angle	2 degrees	0.5	±1 degree
Friction angle	4 degrees	1.7	±2 degrees
Cohesion	0.1 t/m^2	0.05	±0.1 t/m^2
Unit weight	2 t/m^2	0.1	±0.2 t/m^2

Table 5.4: Values for parameters used in probabilistic stability analysis. All parameters have a normal distribution.

As with Ella Landslide, parameters important to the stability of the Amphitheatre Landslide can be assessed using sensitivity analysis based on a deterministic slope stability assessment (plots shown in Appendix IV). The sensitivity of factors considered likely to have an influence on the stability or be variable for different blocks in the landslide complex (failure plane angle, failure plane friction angle, slope height and upper face width) indicates that the model is particularly sensitive to failure plane angle and failure plane friction angle. The factor of safety with the mean parameter values shown in Figure 5.9 and Table 5.4 in a drained condition is 2.1, while considering parameter distribution in a probabilistic analysis, the probability of failure is 2.6 % (Appendix IV).

Hydrological influence

The deep incision of the Ponui, Makara and Te Apiti Streams, and the narrow character of ridges in these catchments means that the regional groundwater table is unlikely to be any where near ridge top level (discussed in Section 4.2.2). It is considered that as the Amphitheatre Landslide is now perched well above stream base level, the area of the landslide complex is unlikely to be affected by the regional groundwater table. In the lower part of the Ponui and Makara Catchments the Makara Formation contains relatively uncemented sand beds that may be tens of centimetres thick (Pettinga, 1980) and could be locations where confined aquifers occur and elevated pore pressure levels may develop. In the upper part of the catchment, however, the only sand beds which occur within the stratigraphy are thin, fine grained and commonly cemented and the permeability of these layers is not dissimilar to the surrounding material. It is subsequently inferred that no pore pressure development on a perched water table was involved in the failure of the Amphitheatre Landslide complex. The tuffaceous horizons often show a degree of "weeping", however, due to the thickness of these (< 20 mm) it is considered very unlikely that any significant pore pressures could develop on them. The main flow of groundwater through the rock mass is hence likely to be defined by secondary permeability along rock mass defects. The secondary permeability defined by dilated defect sets is not thought to relate to the initial failure of intact slopes as rock mass dilation of existing joints and faults is inferred to occur at the time of slope failure.

At the time the block under consideration failed it is inferred that while there may have been some groundwater in the slope, it is unlikely the slope would have been anywhere near saturation levels. The sensitivity of the slope to a range of groundwater levels indicates that the slope may fail when it is approximately 42 % filled (Figure 5.10) and as saturation drops below this level the factor of safety rapidly increases. By the time the saturation level has dropped to 10 % of the slope the factor of safety has increased to 1.95. This low value of about 10 % saturation of the slope is considered to be a realistic value, and a factor of safety of 1.95 will be used to consider the influence of seismic ground motion on slope stability.

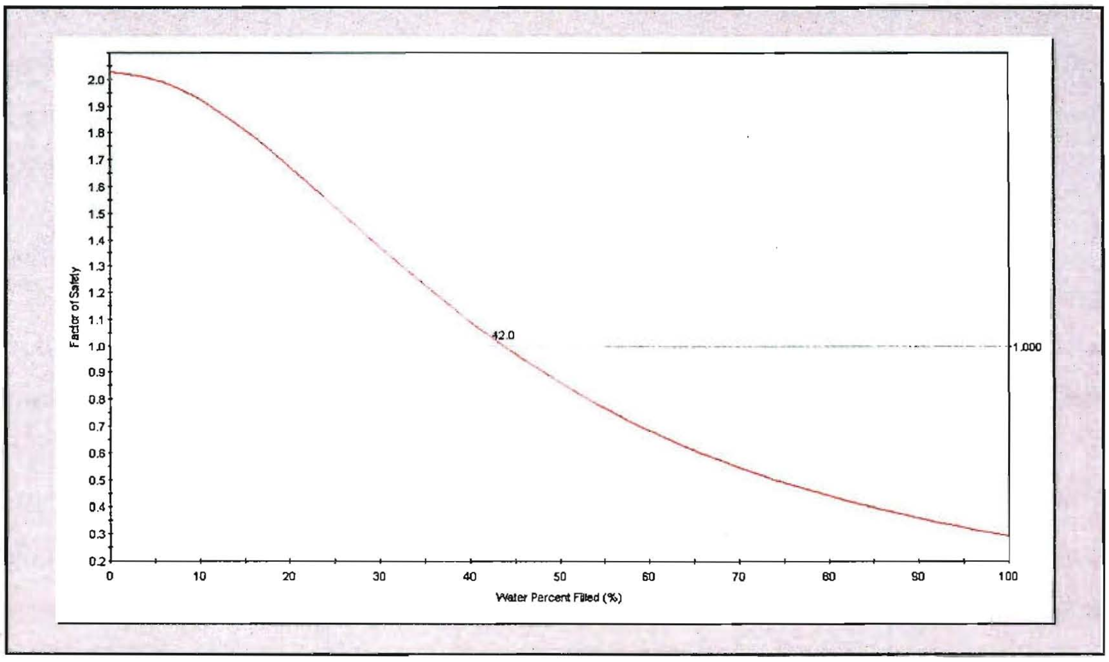

Figure 5.10: Sensitivity of Amphitheatre Landslide to hydrological conditions in a probabilistic stability analysis, using parameters defined in Figure 5.9 and Table 5.4. Plot shows that at 42% saturation the factor of safety is 1.0.

5.3.2 Pseudostatic stability of Amphitheatre Landslide

The occurrence of seismically triggered deep-seated landslides in this catchment (Pettinga, 1980; Pettinga, 1987a), and elsewhere during the 1931 Hawke's Bay earthquake (e.g.

Marshall, 1933) as well as globally (e.g. Keefer, 1984; King et al., 1989) leads to the inference that seismicity may play a dominant role in triggering deep-seated slope failures in the catchments adjacent to the southeastern margin of the Maraetotara Plateau.

It is possible to consider the seismic stability of the Amphitheatre Landslide using the static slope stability model (developed in the previous section) and applying a range of horizontal peak ground acceleration values that represent the influence of earthquake induced strong ground motion on the slope. Using the deterministic method for a pseudostatic analysis, it is predicted that the block would fail under levels of seismic ground motion as low as 0.035g. This low level of seismic ground motion is likely to be exceeded on a regular basis (Stirling et al., 1998), and it is expected that the slopes in the catchments under consideration are not this sensitive to such a low level of seismic perturbation.

Failure of the selected Amphitheatre Landslide block is bracketed between 1952 and 1964 by aerial photograph interpretation, which shows that cracking in the head zone of the block occurred during this period. Also during this interval there were three recorded earthquakes which might be considered as potential triggering mechanisms for this slope failure (Figure 5.11), and using strong ground motion attenuation relationships, the peak ground acceleration at the study site can be inferred.

Attenuation relationships

The relationship between ground motion parameters that decrease with increasing distance from the earthquake epicentre may be termed an attenuation relationship, and these consider the attenuation of parameters such as peak ground acceleration from a defined earthquake source through defined crustal conditions (Kramer, 1996). The use of attenuation relationships (e.g. Matuschka and Davis, 1991; Abrahamson and Silva, 1997; Boore et al., 1997; Campbell, 1997; Sadigh et al., 1997) for assessing the influence of these three earthquake events on the Amphitheatre Landslide site can give an indication of what level of strong ground motion would be realized at the selected site. A comparison of data predicted from these five attenuation relationships is shown in Table 5.5. It is acknowledged that attenuation relationships are being developed which may be more appropriate for the New Zealand situation (e.g. McVerry et al., 1998; McVerry et al., 2000), however, these have not been published in full and therefore cannot be included in this analysis.

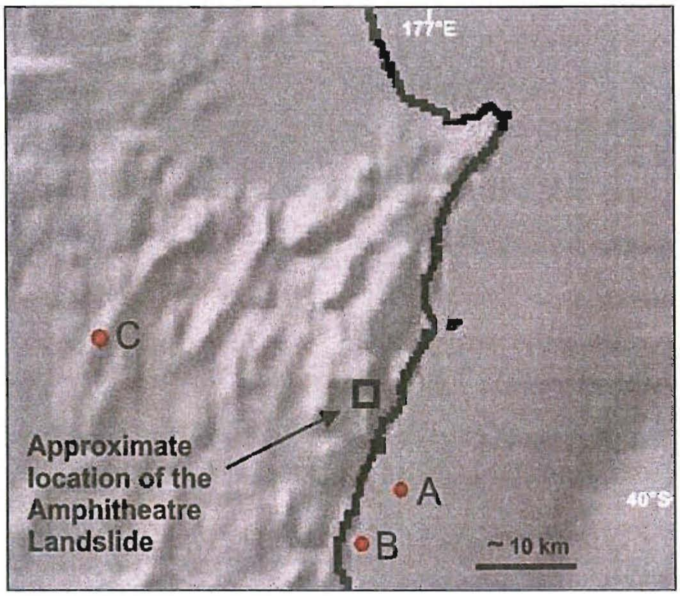

Figure 5.11: Map showing the location of three earthquake epicentres (A, B and C) in relation to the Amphitheatre Landslide. Earthquake A occurred on August 28, 1952 and was M_L 5.8 at 12 km depth, earthquake B occurred on August 29, 1952 and was M_L 5.3 at 12 km depth, and earthquake C occurred on January 31, 1958 and was M_L 6.1 at 12 km depth. The average error for epicentral location for these three events is 12 km E and 7 km N. Map modified from GeoNet database, (http://data.geonet.org.nz/QuakeSearch/index.jsp, accessed on 21 January, 2005).

Date	M_L	Dist. from site (km)	(Boore et al., 1997)	(Abrahamson and Silva, 1997)	(Campbell, 1997)	(Sadigh et al., 1997)	(Matuschka and Davis, 1991)
28/8 1952	5.8	11.5	0.13g	0.42g	0.25g	0.17g	0.23g
29/8 1952	5.3	16	na	na	0.12g	0.09g	na
31/1 1958	6.1	28	0.08g	0.13g	0.1g	0.08g	0.13g

Table 5.5: Comparison of predicted peak ground acceleration at the Amphitheatre Landslide site for three separate earthquakes as predicted by different attenuation relationships. M_L is the Local Magnitude.

Given the data available for the earthquakes and site characteristics the following parameters are assumed: The fault will be reverse if no "other" option is available (no focal mechanisms are available for the events), site materials are variably rock, soft rock and stiff soil, the shear wave velocity for the upper 30 m of soil is 700 m/s (Williams et al., 1997) and the depth to basement is 1.0 km. A limiting assumption of the data used for the attenuation relationships is that the local magnitude documented for these seismic events is equivalent to a moment magnitude.

While the attenuation relationships show that all three of these earthquake events could potentially have provided sufficient ground acceleration to equal the low yield acceleration calculated for the Amphitheatre block by pseudostatic analysis (0.035g), the M_L5.8 event exceeds the others for all five relationships. The attenuation relationships predict an average PGA of 0.24g, however, if the outlier of 0.42g is ignored this lowers to 0.2g which is considered to be a realistic value for an earthquake of this magnitude.

The yield acceleration predicted by pseudostatic analysis would have been exceeded in all three of these events and many times on a time scale of even hundreds of years. There is no evidence of such a high recurrence of deep-seated slope failures and this leads to several possibilities:

- The slope model has some error within it which gives a lower than realistic yield acceleration
- The block modelled is not representative of the development of the Amphitheatre Landslide or other landslides in the study catchments; or,
- The peak ground accelerations predicted by the attenuation relationships used are excessively high for the earthquakes analysed.

We know that the 1931 Hawke's Bay earthquake pre-conditioned the 1976 Ponui Landslide and probably the 1976 Waipoapoa Landslide for failure, and hence must have induced significant strong ground motion in the landscape. This event did not, however, induce massively widespread deep-seated slope failure in all the surrounding catchments as might be expected if 0.035g is a typical yield acceleration for such failures. The possibility that there is a poorly represented parameter in the slope stability model must therefore be considered further.

A reconsideration of slope model parameters

If it is considered that the Amphitheatre Landslide block was at unity (Fs = 1.0) with strong ground motion of 0.2g acting upon it, then back analysis can be used to assess the sensitivity of the slope to variation of specific parameters. One parameter that the slope can be shown to be particularly sensitive to is the shear strength of the failure surface material (Section 5.3.1) which is determined by laboratory testing to have a friction angle between 2° and 5°.

By setting the seismic ground motion to 0.2g and varying material shear strength, the slope is at unity with a friction angle value of 13.5°. A higher friction angle may be more realistic for the residual strength failure surface when compared to published Θ'_R values for pre-sheared failure surfaces in the range of 6.0° – 12.5° (Sugden et al., 1977; Pinckney et al., 1979; Fell et al., 1988) where most values fall between 8° – 10°.

To calibrate the laboratory derived shear strength parameters for the critical stratigraphic horizons in the Makara Formation an attempt was made at back calculating the strength of the nearby Waipoapoa Landslide failure surface as this is inferred to be similar to that tested (Pettinga, 1987b). As this landslide is inferred to have been initiated during the Hawke's Bay earthquake (and fully failed in 1976), attenuation of energy from the 1931 Hawke's Bay earthquake event was modelled to predict a peak ground acceleration at the landslide site. Due to landslide failure model uncertainties, difficulties modelling a slope failing on a surfaces dipping uphill at 15°, and the uncertainty of the strong motion attenuation no useful results were obtained from this exercise.

If a friction angle range of 10° to 13.5° is considered in the static stability model of the Amphitheatre Landslide then the factor of safety for the slope varies between 4.8 and 6.6. These are very high factor of safety values. However, sensitivity analysis shows that the other variable which the model is specifically sensitive to is failure surface angle and this is as low as 2°, so it might be expected that the factor of safety would be high.

5.3.3 Newmark analysis

To carry out a detailed Newmark analysis it is necessary to have a strong ground motion record which represents the earthquake of interest and it was considered that it might be possible to find a record from a strong motion recorder that matches the characteristics of the $M_L = 5.8$ event which occurred in 1958 (source distance, magnitude, site geology). While digitised strong motion records are available for New Zealand earthquakes from 1966, extensive database searches have failed to find a record which is thought to accurately represent this event. For the purposes of this study the simplified Newmark method (Jibson, 1993; Jibson and Jibson, 2003) is therefore adopted.

The average horizontal displacement for the Amphitheatre block in the direction of failure is measured as being 0.4 m. Combined with the yield acceleration, this displacement can be used as input into the Newmark program of Jibson and Jibson (2003) to back analyze the magnitude of the earthquake required to cause this displacement. The Newmark program predicts that an Arias Intensity of 1.45 m/s is required for block displacement and this translates to a magnitude 6.4 earthquake. This exceeds the magnitude of any earthquake in the vicinity at the time, however, if the arbitrary 10 cm Newmark displacement is used (some displacement may be due to post-seismic slide mass deformation), the relationship predicts a magnitude 6.0 earthquake which, while greater than the $M_L = 5.8$ earthquake inferred to have triggered the slope failure, is closer to being a realistic value.

The methods used for dynamic stability modelling have limited value as tools for predicting earthquake magnitudes for prehistoric slopes or slope failures which occurred prior to collection of detailed strong motion records. Even when a specific strong motion record can be correlated to a slope failure there are significant uncertainties involved with strong motion recorder sites, geological and topographical amplification and limitations in the actual representation of ground motion (e.g. Murphy et al., 2002), as well as uncertainties inherent in an static slope stability model. Despite these limitations stability modelling methods can give some indication of how short-term tectonic forcing affects slopes in a specific landscape.

5.3.4 Discussion of deep-seated slope stability in the Hawke's Bay field area

The Amphitheatre Landslide complex is inferred to have been initiated during the late Pleistocene, when the basal failure surface was at stream base level in the catchment headwall, and to have become perched above base level with deepening incision of the Ponui Stream. The proposed mode of failure of the Amphitheatre Landslide complex is by scarp retrogression controlled by the periodic disturbance and failure of large blocks of intact mudstone. A ~3.2 x 10^5 m^3 block is observed to be displaced some 0.4 m down dip and this block is representative of the size, geometry and failure mode of the large blocks of intact mudstone by which scarp retrogression is occurring. Subsequently the stability of the Amphitheatre Landslide complex can be modelled based on the stability of the slope prior to the failure of this block.

Static, pseudostatic and dynamic stability modelling has been used to describe the stability of the pre-failure slope, and ultimately to try and quantitatively determine the trigger for the slope failure. Stability modelling has also been used to assess the sensitivity of the slope to factors such as hydrological conditions, failure plane angle and material strength, and this shows that the failure plane angle and material strength are critical factors in the model.

Stability analysis shows that the selected Amphitheatre block is likely to have required some external perturbation to induce failure. The failure of this block is bracketed between 1952 and 1964 by air photo interpretation, and ground accelerations in the order of 0.2g are predicted from attenuation modelling of three moderate magnitude earthquakes which occurred during this time period. Many parameters affect the propagation of earthquake ground motion (e.g. site foundation materials and source distance) and these introduce uncertainty to empirical attenuation models. There are other aspects to strong ground motion attenuation such as topographic and geologic amplification and modification which may significantly affect the ground motion at a site (e.g. Davis and West, 1973; Shearer and Orcutt, 1987; Geli et al., 1988). Hilltops and ridge crests can show significant amplification for frequencies corresponding to wavelengths which approximate the hill width, while hillsides respond with complex amplification – de-amplification patterns. Paolucci (2002) lists eight instances between 1909 and 1999 where topography has been observed to influence earthquake ground motion and notes the potential for topographic effects to play a role in landslide activation or re-activation.

In the consideration of prehistoric landslides there will always be uncertainty involved with characterisation of an earthquake trigger, however, historical knowledge (e.g. 1931 Hawke's Bay earthquake initiating deep-seated failures in coastal and hill country areas) confirms that large earthquakes do certainly play an important role in slope destabilisation leading to deep-seated failures. The static stability of the Amphitheatre Block indicates that while this landslide is stable under aseismic conditions slopes are likely to be very sensitive to short-term tectonic forcing once critical stratigraphic horizons are exposed. In this specific case, the Amphitheatre Landslide complex is inferred to have been active for tens of thousands of years with lateral and head scarps periodically being pre-conditioned for failure by earthquake activity leading to rock mass dilation of existing discontinuities allowing large blocks of intact mudstone to becoming detached and subsequently degraded. Here the exposure of the tuffaceous critical stratigraphic horizons is a key factor in combination with incident strong ground motion. Similar conditions to those which have allowed the Amphitheatre Landslide to fail are inferred to be involved in the numerous other deep-seated landslides in the three catchments under consideration, as well as catchments developing in the other significant areas of Tertiary soft rock in New Zealand.

5.4 Slope stability modelling summary

The two landslides chosen from the representative soft rock catchments (Ella Landslide in North Canterbury and Amphitheatre Landslide in Southern Hawke's Bay) have been considered in terms of slope stability models based on data from field investigations and laboratory testing. Stability modelling of the two deep-seated soft rock landslides provides an indication of the sensitivity of slopes in bedded Tertiary soft rock sequences to factors such as material strength, failure plane angle, release surfaces, hydrological conditions and seismic ground motion. Under realistic hydrological conditions (slope failures involving ridges are unlikely to be saturated and the occurrence of perched water tables having an influence on slope destabilisation has been discounted in both field areas based on lithological characteristics), it appears that neither of these slopes are likely to have failed under static conditions.

A possible source of an earthquake trigger is inferred for the failures, however, uncertainty involved in certain parameters in the failure models (e.g. headscarp release mechanisms and material strength), and the empirical relationships employed to infer the earthquake trigger limit the certainty of this inference. What is clearly shown by the quantitative slope stability assessment is that under the inferred hydrogeological conditions both of the landslides are statically stable and would be unlikely to fail without some form of external trigger (such as an earthquake). This is supported by numerous documented cases of deep-seated landslides being triggered by earthquake ground motion, while there appears to be a paucity of documented cases of deep-seated failure of undisturbed rock slopes under static conditions in the natural environment.

Chapter Six

6.0 Landscape evolution

Deep-seated bedrock landslides have been shown to be important in the geomorphic development of a variety of landscapes (e.g. Prebble, 1987, 1992; Anderson, 1994; Schmidt and Montgomery, 1995; Burbank et al., 1996; Densmore et al., 1997; Hovius et al., 1997; Hovius et al., 2000; Korup, in press; Roering et al., in press). The majority of landscapes where the influence of deep-seated landslides is documented are mountainous areas with high and steep relief, such as the Southern Alps of New Zealand. The regions of New Zealand underlain predominantly by fine grained Tertiary soft rock strata are not mountainous, however, but are generally characterised topographically by low to moderate relief hill country. The role of bedrock landslides as an erosional process in the evolution of Tertiary soft rock landscapes is considered to be significant, if not dominant. An understanding of the controls on deep-seated landslide occurrence is critical to understanding the evolution of these landscapes and the sensitivity of landscape development to the roles of tectonic and climatic forcing.

The representation of geomorphic processes in numerical landscape evolution modelling of mountain range scale terrain has progressed significantly in the last 15 years. Numerical landscape evolution models have developed from one dimensional models considering geomorphic processes by a single diffusion coefficient (e.g. King et al., 1988; Stein et al., 1988), to three dimensional models such as the finite difference model ZSCAPE (Densmore et al., 1998; Ellis et al., 1999) which allows for regolith production and diffusion, fluvial incision and sediment transport, and bedrock landsliding. ZSCAPE is thought to be the first numerical landscape evolution model to incorporate bedrock landslides, whose occurrence is considered as strongly stochastic with discrete events controlled by base level fall as a hillslope toes becomes steeper, less buttressed and more prone to failure. Landslides are modelled to occur on failure planes passing through the toe of the hillslope and trigger when shear stress in the slope becomes equivalent to shear strength as defined by a rock mass strength dependant critical hillslope height.

In landscapes where the spatial occurrence and geometry of deep-seated landslides is fundamentally controlled by rock mass defects, and landslide failure planes are pre-defined by thin weak stratigraphic layers, this slope height threshold approach does not describe the occurrence of this style of slope failure. It is a main hypothesis of this project that deep-seated landslides in Tertiary soft rock terrain are strongly defect controlled, and inherently non-stochastic. By quantifying controls on the geometry, distribution and initiation of defect controlled deep-seated landslides it should be possible to clearly define why, how and where they occur in time and space and this in turn could allow for the development of landscape evolution models that are more representative of landscape development in areas where rock mass properties control deep-seated mass wasting processes.

New Zealand contains significant areas underlain by fine grained marine Tertiary soft rock successions (approximately 35% of the North Island and 10% of the South Island, refer Figure 1.1) and many of the landscapes which develop in these successions can be considered in terms of specific controls and characteristics which include:

- Tectonically active environments controlling landscape uplift and the occurrence of periodic seismicity
- Deeply incised stream networks driven by long-term tectonic and climatic forcing
- Lithology characterised by weak rock material and well developed rock mass defects; and
- Ubiquitous and widespread slope instability.

The way in which factors such as these are controlling landscape development in the field sites chosen for this project (in Southern Hawke's Bay and North Canterbury) defines a template for the controls on landscape evolution which are broadly applicable to other areas of Tertiary soft rock terrain.

The Southern Hawke's Bay field site covers three small to moderate sized coastal catchments (in the order of 15 – 20 km^2 planimetric area, note that the Makara Catchment is actually a sub-catchment of the Tukituki Catchment and only the upper catchment is considered) with significant topographic relief over ~500 m elevation. The geomorphic development of these selected catchments is considered to be fundamentally related to the large proportion of the landscape affected by deep-seated (bedrock) slope failure. The North Canterbury field site is also a small coastal catchment (~4 km^2) which contains comparatively few instances of deep-seated bedrock slope failure, however, the impact of isolated deep-seated slope failures on the geomorphic evolution of this catchment is significant and affects the landscape for thousands to tens of thousands of years.

The main focus of this chapter is to consider the controls on landscape development at the southeastern margin of the Maraetotara Plateau, Hawke's Bay. The data collected during this study can be assimilated with other available data to quantify controls on deep-seated slope failure and subsequently define a framework for a landscape evolution model. The North Canterbury example is considered to demonstrate how isolated instances of deep-seated slope failure can also significantly affect catchment development.

6.1 Catchment evolution on the southeastern margin of the Maraetotara Plateau, Hawke's Bay

The Makara Basin sedimentary succession comprises the alternating sandstone and mudstone flysch of the Miocene age Makara Formation overlain by the erosion resistant Pliocene age Te Aute limestone, the latter defining the Maraetotara Plateau. The catchments developing on the southeastern margin of the Maraetotara Plateau provide an excellent case study for the development of a landscape with the controls and characteristics as outlined in the previous section and one in which rock mass defect controlled deep-seated landslides are proposed to have a dominant influence on landscape evolution and catchment morphology.

6.1.1 Style of rock mass defect controlled deep-seated landsliding within the Makara Basin sedimentary succession

Within the Makara Basin deep-seated landslides are widespread, and the style of failure and failure mechanisms of these vary within the structure of the Atua Syncline (see Figure 2.12). Deep-seated landslides in this study area are failing on bedding horizons, identified in this study as *critical stratigraphic horizons* and have release geometries controlled by rock mass defects (intersecting joint and fault sets), and stability analysis and historic examples have highlighted the importance of earthquake triggering in the initial movements of landslide block failures. The failure styles can be subdivided into three distinct groups defined by the dip of the strata (low angle retrogressive block slides, moderate angle retrogressive block slides and wedge failures) and each of these three styles will be considered individually. Figure 6.1 indicates the spatial relationship of deep-seated landslides to the structure and stratigraphy and how these appear in the landscape.

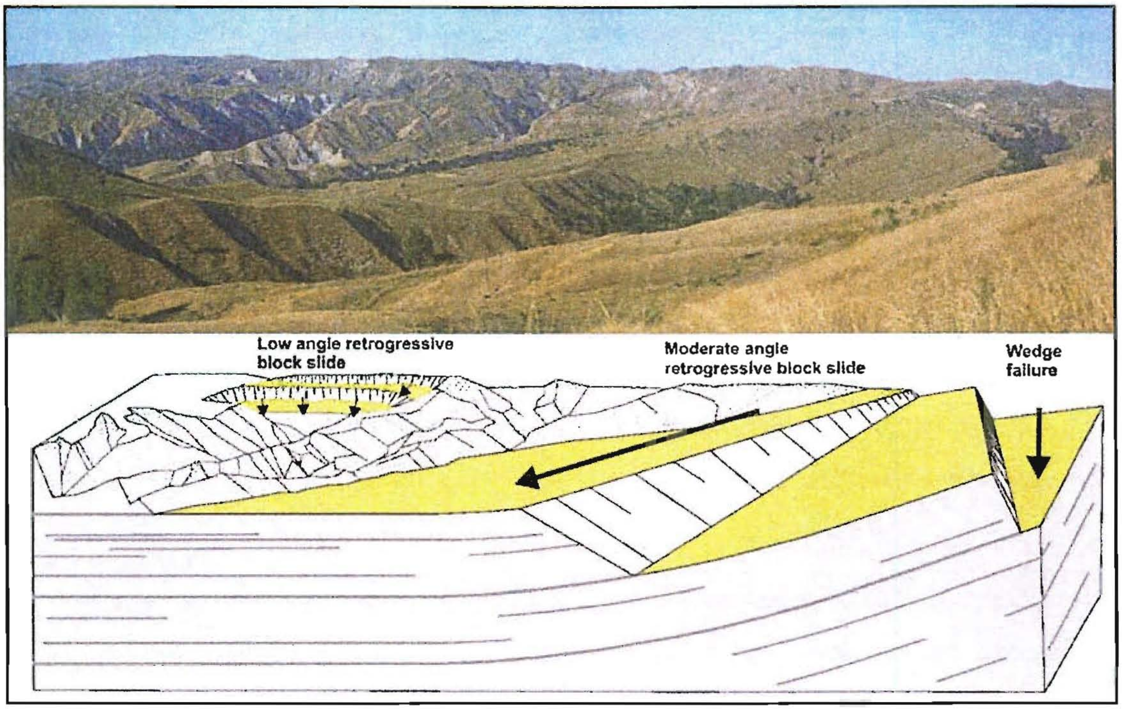

Figure 6.1: Photograph and interpretive block diagram showing variation of landslide failure type with location within the east limb of the Atua Syncline. The three failure styles recognised are indicated showing general direction of failure. Photograph courtesy of J. Pettinga.

Low angle retrogressive block slides

Where dips are less than approximately 5 – 7° slopes will fail (either up, down or obliquely across dip) as low angle block slides. The Amphitheatre Landslide is a prime example of this (see Figure 3.11 and Figure 4.6) and is the example indicated in Figure 6.1. Rather than the complete slide mass failing in one event these are landslide complexes that will retrogress in piecemeal style, with scarp regression being periodically active but debris transport across the failure surface persisting as seasonally enhanced creeping debris flows. Low angle retrogressive block slides are represented in the landscape as essentially flat benches of varying extent. Where low angle retrogressive block slides are evolving by up-dip failure, landslides are represented in the landscape by planimetrically small landslide footprints (such as the landslides which occur in the Te Apiti catchment head, Figure 3.12). When low angle retrogressive block slides are evolving by failing down-dip (e.g. Amphitheatre Landslide) then the footprint of the landslide is significantly larger. This is inferred to be related to factors such failure activity rates relative to stream incision into bedrock. A retrogressive landslide complex occurring on a basal failure surface dipping into the slope is expected to have increased scarp stability and a lower level of activity than in the situation where the basal failure surface dips out of the slope, so consequently if the rate of stream head incision occurs at a comparable rate, less of the landslide footprint is expected to be preserved.

Moderate angle retrogressive block slides

As dips increase to between approximately 7° and 15° landslide failure mode is still by periodic retrogressive scarp retreat involving block slides, however, the representation in the landscape will be as extensive benches with a gradient coincident with stratigraphic orientation. The extensive bench in the centre of Figure 6.1 is a good example of this, as is the Makara Landslide at the head of the Makara Catchment which is approximately 1 km^2 in area.

Wedge failures

As bedding dips increase above approximately 15°, deep-seated landslides are likely to occur as wedge failures defined by bedding plane failure surfaces and intersecting conjugate joint sets. The slide mass is now prone to fail obliquely across dip in a single event, and the Ponui Landslide shown in Figure 6.2 is documented to have failed in this way (Pettinga, 1987a).

Figure 6.2: Ponui Landslide after failure showing the landslide dam formed by the slide debris (right of photo with pine tree plantation running upslope), bedding dips right to left at between 12 and 36° (Pettinga, 1980). The lateral scarp from another wedge failure is evident at the left of the photo covered in native forest. This scarp is just visible at the right hand edge of the photograph in Figure 6.1. Photograph courtesy of J. Pettinga.

The Ponui Landslide is just one in a series of wedge failures occurring in the same area. The Ponui Landslide is failing on a bedding coincident thrust shear (Pettinga, 1987a) and while this instance of a thrust shear defined failure surfaces may be an isolated one (other wedge failures be failing on tuffaceous horizons), it is noted that areas where dips occur at steep enough angles to allow the development of wedge failures are adjacent to areas of thrust faulting that bound the Atua Syncline (Figure 2.12). It might be tentatively inferred that the occurrence of such large wedge failures occurring in single events relates to the development of pervasive bedding parallel thrust shears. No field investigations have been undertaken with this inference in mind and significant further investigation would be required to substantiate this hypothesis.

6.1.2 Fundamental controls on landscape evolution

To define how a landscape is developing in response to a component of rock mass defect controlled deep-seated landslide activity, several specific parameters need to be quantified. Factors such as uplift rates, lithological variation, structure and spatial/temporal controls on the initiation of landslides will need to be numerically defined. At the landslide initiation level there will need to be quantification of incision rates, failure surface location within the stratigraphy, volume and depth of failures, hydrologic variability, and strength properties of the key stratigraphic materials.

The main factors which control catchment evolution adjacent to the Maraetotara Plateau are: long-term tectonic and climatic forcing, lithological variation, structural characteristics, and seismic and climatic events (short-term tectonic and climatic forcing).

Stream network development – the role of long-term tectonic and climatic forcing

Long-term tectonic forcing influences landscape development by uplift, tilting and folding the rock mass, while long-term climatic forcing influences the landscape through the fluctuation of relative sea-level (see Section 1.3). The deeply incised stream catchments and steep valley walls which flank the southeastern margins of the Maraetotara Plateau are a direct reflection of a combination of late Cenozoic tectonic uplift of the Hikurangi Margin frontal wedge, and Quaternary sea-level variation due to orbitally forced glacial and interglacial episodes.

Deep stream incision exposes critical stratigraphic horizons within the stratigraphic section, which define failure planes for deep-seated landslide blocks which area released on intersecting defect sets. As episodes of relative sea-level change occur, and stream base level variation drives periods of accelerated stream incision, some areas of the landscape get decoupled from the current fluvial system. The difference in the land-forms and slope processes acting in these "relict" landscape compared to the "rejuvenated" landscape (see Chapter 2) has a significant influence on catchment morphology. While in relict areas little mass wasting occurs and "abandoned" slopes become quasi-stable, in actively incising areas deep-seated slope failures initiate and remain active while shallow slope failure and slake degradation influence the retreat of steep stream valley walls.

Fluvial processes play an integral role in continued landslide activity as they remove landslide debris as it is delivered to the fluvial system. For large volume landslides in which the slide mass fails in one motion (e.g. the ~$2x10^6$ m^3 Ponui Landslide in 1976), very large amounts of material are introduced to the fluvial system at one time and debris may take years to thousands of years to be removed. These styles of failure commonly dam stream channels and lead to the accumulation of large volumes of sediment, and an example of this is the ~$5x10^8$ m^3 Sunworth Landslide complex which dammed a sub-catchment of the Waipaoa Catchment, Gisborne (Pere, 2003) and caused the accumulation of some 120 Mt of sediment over ~13kyr. In contrast, retrogressive failures that are perched above stream base level introduce debris to the system in much smaller and more gradual quantities. The Amphitheatre Landslide case study shows how blocks of intact material detached from the head scarp eventually degrade into debris and become incorporated in the creeping debris flow active on the basal failure surface (discussed in Chapter 4). The rate at which sediment is introduced into the fluvial channel network by debris flows is in part seasonally dependant, however, observations during the month of February (October to February are the lowest mean rainfall months) showed that the debris flow continued to be active, providing sediment volumes in the order of one to several m^3/day. The removal of such material is obviously important to avoid clogging of the fluvial system and field observation indicates that despite the ephemeral nature of the streams in the upper catchment heads, even low flow conditions in the summer months are sufficient to remove much of the material introduced into the system. As most of the landslides in the study area are propagating by sector instability and retrogressive failure along the headwall, the constant removal of material is a very important aspect of ongoing landslide activity.

Lithology – Critical stratigraphic horizons

There are two units of interest with respect to the evolution of the catchments at the southeastern margin of the Maraetotara Plateau. The erosion resistant Te Aute Limestone controls the extent of the Plateau itself by forming a surficial cap on top of the easily eroded Makara Formation flysch (alternating sandstone mudstone units).

The Makara Formation contains several thin tuff horizons which act as failure surfaces for the widespread deep-seated landslides. These layers are of low strength ($\Theta'_R = 2 - 5°$), appear to be regularly spaced, are often pre-sheared and are laterally continuous (possibly across the full extent of the ~20 x 30 km Makara Basin). The strength, continuity and spacing of these tuffaceous horizons are critical to the evolution and morphology of hillslopes in this area, and it is in this sense that these are considered to be "critical" stratigraphic horizons, i.e. within the stratigraphy these are critical to the occurrence of deep-seated slope failure and the evolution of the landscape. While the intact strength of the Makara Formation may influence rates of stream incision into bedrock, it is not thought to exert any influence on deep-seated slope stability.

The influence of the Te Aute Formation on landslide complex stability and activity has not been investigated in this study, however, it is clear from the relationship of the essentially flat Maraetotara Plateau to the oversteepened upper catchment adjacent to it that the erosion resistant cap defined by this formation retards catchment head progression to some degree, as all the major drainages form steep catchment head-walls against the plateau. Te Aute Formation involved in the Waipoapoa Landslide failure was weathered and broken up (Pettinga, 1980), and Te Aute Formation material involved in the Makara Landslide is similarly observed to be highly discontinuous. While it is difficult to quantify the influence of a cap of intact Te Aute Formation on slope stability without further investigation, it is inferred that the rock mass of this material would be of similar strength to the Makara Formation rock mass and so would not affect slope stability significantly with respect to deep seated failure, and it is the resistance to weathering and other forms of mass wasting that lead it to have such a significant influence on landscape morphology.

Correlation of landslides occurring on the same tuff horizon

It is possible in some instances to correlate deep-seated landslides across the Te Apiti, Ponui and Makara catchments that are failing on the same critical stratigraphic horizon (Figure 6.3).

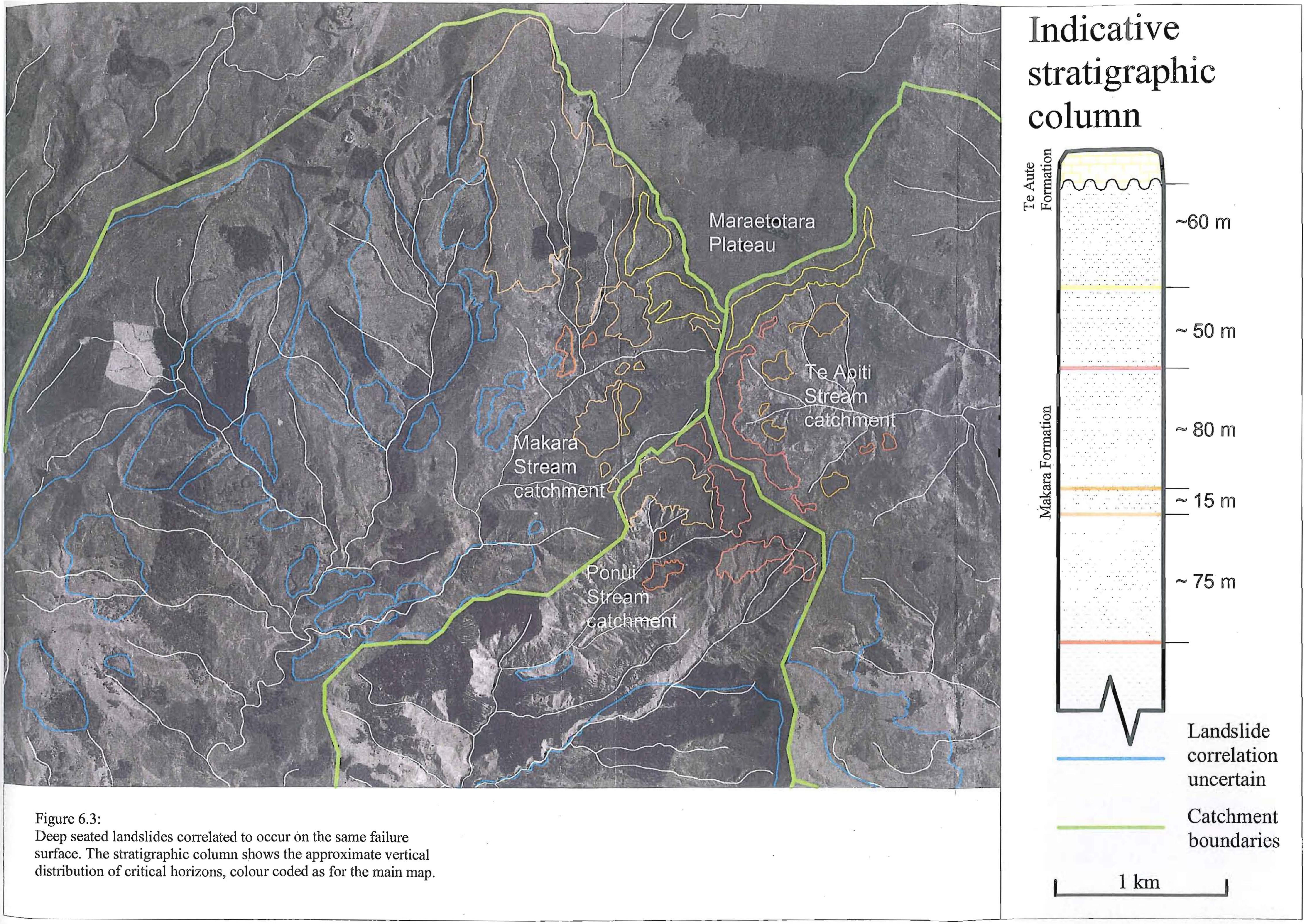

Figure 6.3:
Deep seated landslides correlated to occur on the same failure surface. The stratigraphic column shows the approximate vertical distribution of critical horizons, colour coded as for the main map.

While this can be done with a reasonable degree of confidence parallel to the structural trend (NE – SW), it has a much larger degree of uncertainty when correlation is attempted across frequently occurring normal faults, primarily due to uncertainty in the amount of vertical offset across them. The correlation of widespread deep-seated slope failure occurring on the same stratigraphic horizon highlights the criticality of these horizons.

Structural controls

Landslide geometry is defined by a combination of critical stratigraphic horizon orientation, controlling the basal failure surface, and conjugate joint set orientation defining block release surfaces. Normal fault sets also play a role in landslide geometry, however, and may offset the basal failure surface of deep-seated landslides. The Makara Formation has low dips in the study area which only increase significantly (above ~20°) near the bounding thrust faults. The broad and symmetric nature of the Atua Syncline allows for stratigraphic projections underneath the catchment landscape, and makes the catchment amenable to landscape evolution modelling, as it introduces little in the way of structural complication.

A predominant set of northeast trending normal faults dissect the Makara Basin (discussed in Chapter 2). In the area of the Maraetotara Plateau development of these faults, which define horst and graben features disrupting the generally flat lying surface of the Te Aute Limestone, is attributed to bending moment faulting due to broad arching coincident with the loci of maximum uplift (Pettinga, 2004). To the east, adjacent to the Maraetotara Plateau, normal fault sets relate to extension in the headscarp zone of a massive eastward gravitational collapse. While the major faults defining the escarpment at the eastern edge of the Maraetotara Plateau (see Figure 2.12) have a significant offset (several tens of metres), most minor faults and (NW dipping) conjugate pairs have small offsets in the order of a few to several metres. In the Amphitheatre Landslide vertical offset of the basal failure surface by normal faulting of approximately 1-3 m (refer Figure 4.6) does not appear to inhibit the propagation of slope failure or the development of the landslide complex. Significant offset of the stratigraphy (several to tens of metres) would disrupt the stratigraphically controlled failure plane of deep-seated landslide complexes to the extent that landslide activity might be inhibited and hence such faults may define the extent of certain landslide complexes, however, no clear field evidence for this has been discerned from this study.

Landslide trigger mechanisms – the role of short-term tectonic and climatic forcing

The Amphitheatre Landslide complex is considered to be representative of a wider population of deep-seated retrogressive landslides within the Hawke's Bay study area. The stability of the Amphitheatre Landslide has been considered in terms of the failure of a block of intact mudstone that is inferred to represent the mode of development for the landslide complex (refer Chapter 4). It can be shown using computer based stability modelling that this representative landslide block is stable under static conditions, and is subsequently inferred to have failed following a seismic event. The largest magnitude earthquake to occur in this region in historical times (1931 Hawke's Bay Earthquake) triggered and/or reactivated at least one deep-seated landslide in the study area and several deep-seated landslides near the coast (Marshall, 1933; Pettinga, 1987a). It is inferred from stability modelling of the representative landslide, and from slope failure during this historical earthquake event that the initial triggering of deep-seated landslides in the study area is primarily controlled by seismic events. No documentation of large deep-seated landslides being clearly triggered by the sole influence of climatic events in this area has been located.

To consider landscape development over geomorphic time, the recurrence of seismic landslide triggering events, and the critical aspect of earthquake ground motion accelerations generated by these earthquakes are required. Stirling et al. (2002) predict peak ground accelerations in the study area of 0.3g – 0.4g with a 150 year return period, 0.4g – 0.5g with a 475 year return period, and 0.5g – 0.6g with a 1000 year return period based on a seismic hazard model developed for the whole of New Zealand. Large subduction thrust earthquakes in Hawke's Bay have a calculated recurrence interval of 250 – 400 years calculated for an $M_W = 7.7$ earthquake based on seismological, geodetic and geologic data (Reyners, 2000). Attenuation relationships can be used to infer the level of ground motion (in terms of peak ground acceleration) that might be expected in the study area during such an event. Based on the potential fault rupture plane inferred by Reyners (2000) and using a range of attenuation relationships (Matuschka and Davis, 1991; Abrahamson and Silva, 1997; Boore et al., 1997; Campbell, 1997; Sadigh et al., 1997) for a $M_W = 7.7$ earthquake on a NW dipping subduction thrust fault, the peak ground acceleration in the area of the Maraetotara Plateau can be estimated. Considering the site as a 1 km thick layer of soft rock overlying basement bedrock the attenuation relationships predict a range of peak ground acceleration values of between 0.74g and 1.4g with three of the predictions between 0.74g and 0.89g. The 1.4g prediction is anomalously high and if this is discounted an average peak ground acceleration of 0.8 can be inferred. This value is exceptionally large, when compared to estimates of Stirling et al. (2002), and this reflects the location of the inferred subduction interface rupture plane at shallow mid-lower crustal depth, directly beneath the field area.

Mode of slope failure with an earthquake trigger

It is expected that an earthquake would not cause a slope to totally fail on a deep-seated basal shear plane, but would rather pre-condition the rock mass for failure by dilation of existing rock mass defects and minor displacement on basal shear planes. The Makara Formation rock mass can be considered as a fully discontinuous medium defined by conjugate sets of joints and faults and very weak bedding parallel surfaces. Rock mass defects are typically closed at depth, but the effect of strong ground motion on a rock mass deeply dissected by topographic incision is to cause partial rock mass dilation, particularly on conjugate joint and fault sets. This weakens the rock mass and creates a high degree of secondary permeability, establishing conditions in which water can infiltrate rapidly to critical horizons and develop elevated pore pressures.

Annual to decadal climatic variability plays the dominant role in initiating shallow slope failure, and so controls a significant proportion of sediment production in many soft rock catchments. The role that short-term climatic forcing plays in the deep-seated landsliding process may be more complex. Although the initial landslide trigger is shown to be related to seismic activity, the final failure of a given landslide is thought to relate to decadal cycles of climatic variability and time dependent rock material degradation.

In February 1976 the Waipoapoa landslide failed following almost 374 mm of precipitation in 24 hours (Pettinga, 1987b) (Figure 6.4), and the Ponui Landslide failed some seven months later during a month of similar rainfall (Pettinga, 1987a). It is not thought that a period of high rainfall alone provided the trigger mechanism for these failures (i.e. solely as a function of reaching a pore pressure threshold) as the precipitation level at failure had been exceeded many times since the 1931 Hawke's Bay earthquake. In Figure 6.4 there are two prolonged periods of below average precipitation evident, around 1973 and 1975 and it is thought that these extended dry periods will allow enhanced shrinkage and associated fracturing of clay rich materials. This rock mass dilation results in an increase in the secondary permeability of the rock mass, allowing pore pressure to build up rapidly on defined basal failure surfaces where previously the very low (~10^{-8} m/s) primary permeability precluded this.

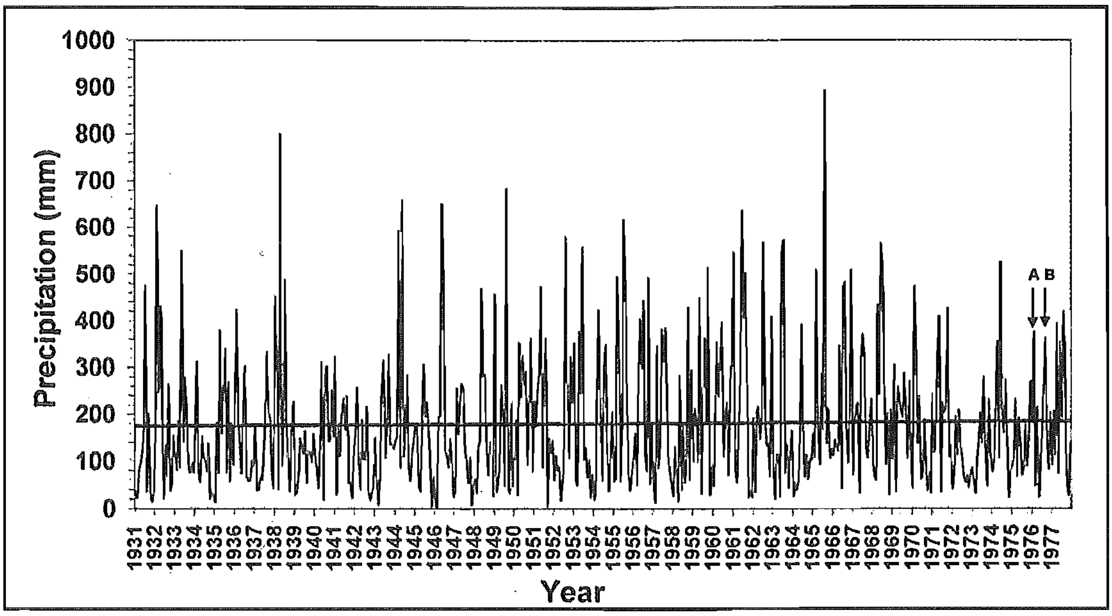

Figure 6.4: Rainfall records for Maraetotara Plateau recorded at Anawai (6142150N 2841800E NZMG 260 series map sheet V22). The horizontal line at 175 mm indicates the monthly average for all 46 years. The two arrows show the timing of the Waipoapoa Landslide (arrow A) and the Ponui Landslide (Arrow B). Data from Pettinga (1980) and Hawke's Bay Regional Council (2004).

It is unlikely to be a coincidence that forty five years after earthquake triggering, the Ponui Landslide failed within seven months of the Waipoapoa event. There are two likely reasons for these coincident failures; the specific climatic conditions with the combination of a long dry period followed by a large precipitation event; and/or, a time dependant material degradation threshold that has weakened the rock material (and specifically the material associated with the failure surface) to reach critical conditions for slope instability. It is likely that both of these factors played a role in destabilising the already defined slide mass of these two landslides, and it is clear that the trigger of the final slope failure relates precipitation induced pore pressure build up. If the failure of the Ponui and Waipoapoa landslides is considered in a context of geomorphic time (10^3 – 10^5 yrs) then the entire process from earthquake driven "pre-conditioning" of the rock mass by dilation to the full failure of the slide mass some 45 years later, can essentially be considered as one event and in terms of landscape evolution the slope failure could be thought of as co-seismic.

In the two examples discussed (Ponui and Waipoapoa Landslides) the slide mass has failed in one large motion, however, many landslides within the study area are thought to be failing in a retrogressive fashion by initiating and remaining active through the influence of earthquake ground motion acting on slopes, and where failure geometry is defined by critical stratigraphic horizons and conjugate joint sets. The role of short-term tectonic and climatic forcing in triggering deep-seated slope failure is complex, and while this study is able to show that deep-seated landslides fail due to a combination of earthquake shaking and precipitation patterns, which is considered adequate in terms of the triggering mechanism for deep-seated slope failure over geomorphic time, the mechanics and details of how such large masses of intact rock are able to become separated from a slope warrants further investigation.

6.1.3 Landslide initiation and persistence: The maintenance of a stepped landscape morphology

Catchment slopes will become unstable once stream incision exposes critical stratigraphic horizons and deep-seated failure will initiate and develop while the failure surface is still near stream level. Figure 6.5 shows an example of a deep-seated slope failure with the basal failure plane at stream level. An exposure of an in-situ critical stratigraphic horizon could be visually correlated to the basal failure surface of this landslide. With the continued deepening of valleys by stream incision, landslide complexes will become progressively perched in the landscape above stream base level. Although these landslides are no longer directly connected to the current base-level they will continue to actively retrogress. Based on field observations and analysis it is clear that once initiated many of these failures will stay active for tens of thousands of years, until the critical stratigraphic horizon on which they are failing is obliterated by stratigraphically lower failures, i.e. they effectively pass out of the catchment slopes by continued lowering of the landscape.

Figure 6.6 shows the current slope configuration between the Te Apiti and Ponui catchments and some predicted slope configurations at undetermined time intervals into the future. These time intervals directly relate to accelerated base level lowering events and it is difficult to determine the temporal occurrence of these. As landslides occur on stratigraphically lower surfaces, failures on higher surfaces will be further destabilised and will retrogress towards the common ridge top. At a certain point the upper level critical horizons will be removed out of the catchment slopes through the ridge top. At present there is a mantle of Te Aute Formation derived limestone colluvium on the ridge top between the Ponui and Te Apiti catchments that is a remnant of the erosion resistant Te Aute Formation limestone cap-rock defining the Maraetotara Plateau to the north.

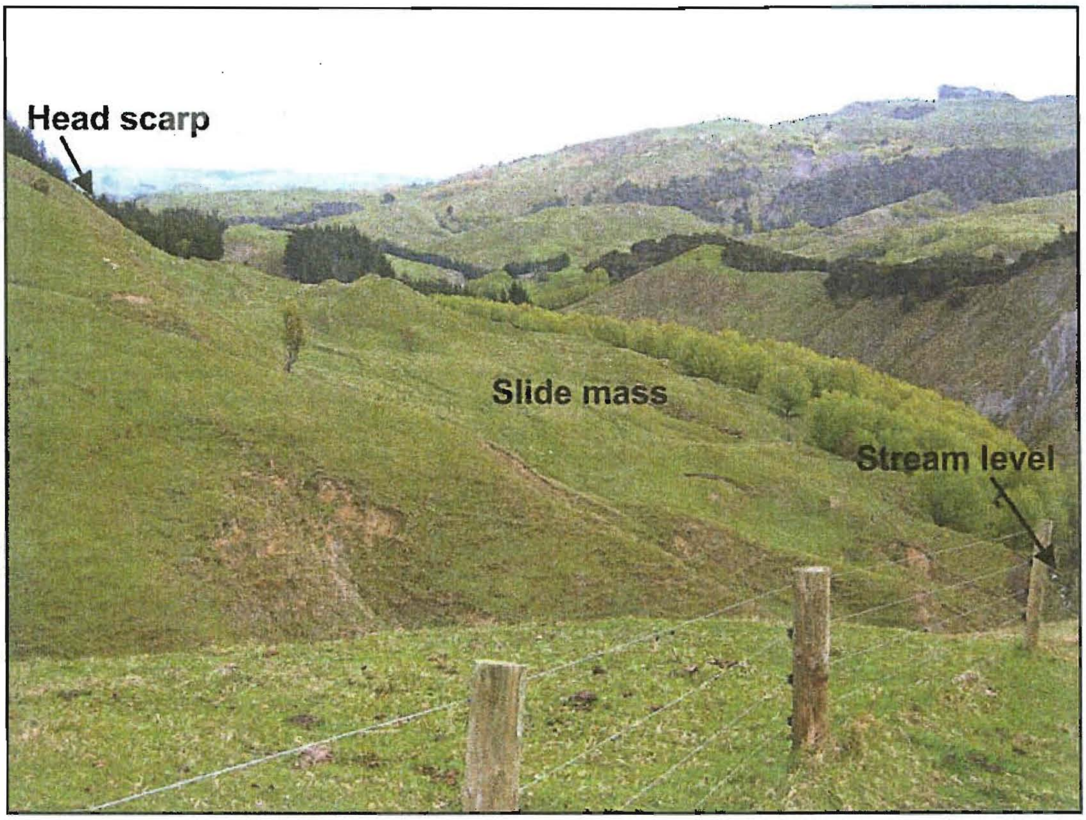

Figure 6.5: Deep-seated landslide with the basal failure plane at the level of Makara Steam. Photograph taken looking southwest from 6137750N 2843500E (NZMG 260 series map sheet V22).

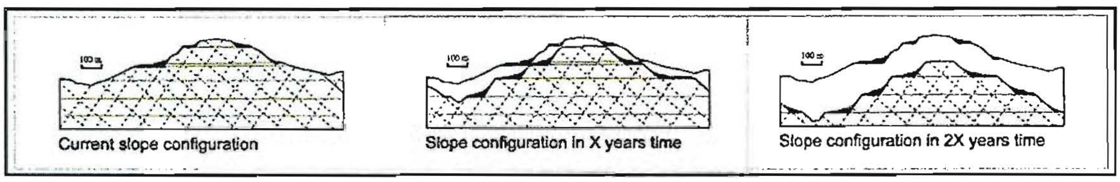

Figure 6.6: Current and future predicted cross sections between the Ponui and Te Apiti catchments. Based on cross section B – B' in Figure 3.12.

The stepped landscape morphology

It is readily apparent that slopes developing in this manner obtain and maintain a distinctive, stepped profile. This is defined by the "treads" of the steps which are coincident with critical stratigraphic horizons and the "risers" which are defined by the stratigraphic spacing between these (Figure 6.7).

The stepped profile will persist throughout the lifespan of catchment evolution within the Makara Basin sedimentary succession so long as tuff layers occur throughout the sequence. The establishment and maintenance of the stepped profile is dependant on the characteristics of the Makara Formation lithology within which it is developing (primarily rock mass defects) and so this landscape morphology will only be maintained while catchments are developing in the Makara Formation. The Mid Cretaceous to Oligocene lithologies which underlie the Makara Formation are significantly more deformed and exhibit a significantly different style of deep-seated slope failure (Pettinga, 1980, 1992). Subsequently, the morphology of the landscape will be drastically altered once these sequences are introduced into the study catchments and it is unlikely that the stepped landscape morphology will be maintained.

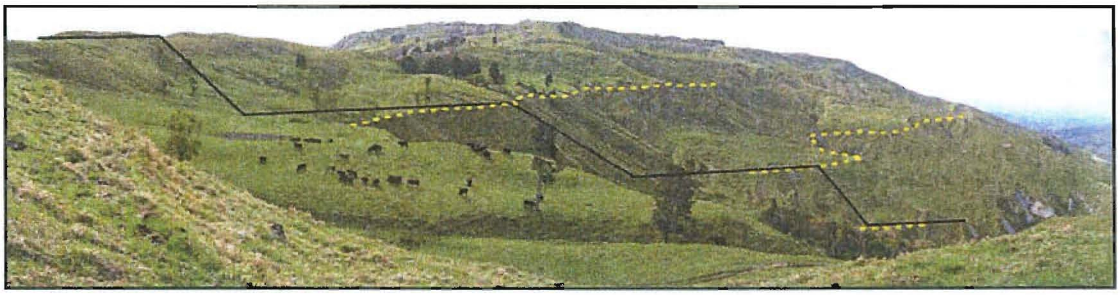

Figure 6.7: Stepped landscape morphology in the head of the Te Apiti catchment. The dashed yellow lines indicate the approximate location of critical stratigraphic horizons defining the basal failure surfaces of extensive areas of deep-seated slope failure. Photograph taken looking north from 6138150N 2844250E (NZMG 260 series map sheet V22).

A core concept in the consideration of landscape evolution and tectonic geomorphology is the "steady state" approach to the development of landscapes over time (Burbank and Anderson, 2001). This considers that the combination of tectonic uplift and hillslope denudation results in a specific landscape morphology, which is subsequently maintained over time with summit heights, valley wall steepness and topographic relief fluctuating around long-term average values. This implies that uplift rates are to some extent balanced by catchment hill slope denudation rates. The predicted maintenance of the stepped landscape morphology shown in Figure 6.6 infers that in the catchments developing at the southeastern margin of the Maraetotara Plateau a steady state of some form prevails. It is inferred that the maintenance of this steady state development of the landscape's stepped morphology will continue, while catchments are developing in the Makara Formation alternating sandstones and mudstones, and that the increments of vertical slope lowering are controlled by the spacing of the critical stratigraphic horizons. While vertical landscape lowering is directly influenced by the rate of stream incision into bedrock, no connection between the rates of these two processes is implied.

Evolution of catchment heads adjacent to the Maraetotara Plateau

For all of the Te Apiti, Ponui and Makara streams, catchment head development is clearly defined by the ongoing retrogression of deep-seated landslide complexes (refer Figure 6.3). The aerial extent of deep-seated landsliding in the Te Apiti catchment head is limited by the orientation of strata, as landslides are mostly failing in a slightly up-dip direction. Conversely, in the Makara catchment head extensive retrogression of the Makara Landslide defines a large planimetric area, as bedding dips out of the slope at approximately 10°. The planimetric area of retrogressive landslide complexes is controlled by a balance between the rate of head scarp regression of the landslide complex, and the rate of headward propagation of the incising stream gully system as it obliterated the landslide footprint. Rates of river incision into bedrock have been found to be controlled by factors such as stream power, rock strength, sediment supply and grain size (Sklar and Dietrich, 2001), however in the catchments under consideration in this study, intersecting rock mass defects (joints and faults) and rock material degradation are inferred to be the primary controls on catchment incision and slope parallel enlargement (refer Section 4.2.2).

A clear limiting factor in the progression of catchment heads is the presence of the erosion resistant cap of the Maraetotara Plateau, and catchment head incision rates are likely to be significantly limited where intact Te Aute Formation is involved. The Maraetotara Plateau essentially represents a remnant of the very early stages of landscape evolution from the time of emergence that has had a continuing influence on the rate of overall hillslope lowering. It is likely that when this cap is eroded out of the landscape the rate of catchment development will be significantly impacted which may have flow-on effects on factors such as sediment flux volumes.

Sediment contribution from deep-seated landslides

While no estimate to quantify the daily or annual sediment input from any deep-seated landslide has been attempted in this study, observation at the Amphitheatre Landslide during dry conditions indicates sediment contribution from this source point is in the order of m^3 day^{-1}. In wet conditions the sediment contribution would be higher as seasonal variations in the activity of the creeping debris flow on the basal failure surface is well established (Pettinga, 1980).

The 1976 Ponui Landslide dam reservoir had an estimated volume in the order of 3 x 10^6 m^3 (de Leon, 1977, cited in Pettinga 1987a) and this was essentially at full capacity within 25 years of dam creation. As a crude estimate this equates to a daily sediment flux of 330 m^3, or as the contributing area is in the order of 8.6 km^2 (± 4%) a lowering rate of some 14 mm/year. While it is not possible to differentiate the contribution of deep vs shallow landslides to this sediment flux, it is clear that as a significant proportion of the upper catchment is subject to deep-seated failure these will be a significant contributor. If the average contribution from the Amphitheatre landslide is considered at 4 m^3 day^{-1}, this has an estimated lowering rate of 106 mm/year which far exceeds the aerial sediment production average calculated from sediment accumulation behind the Ponui landslide dam, and even at 1 m^3 day^{-1} sediment production from the Amphitheatre Landslide would be double that of the average.

There are many other deep-seated landslides contributing sediment to the Ponui Catchment flux and, while differentiating sediment production mechanisms for catchment sediment flux is not a focus of this project, it is apparent that the contribution of deep-seated landslides is significant. The contributors to the Ponui dam infilling event are a small aspect of this, the Ponui Landslide itself had an estimated volume of 2.5 x 10^6 m^3 (Pettinga, 1987a) and the Waipoapoa Landslide some 8.6 x 10^6 m^3 (Pettinga, 1987b). Much of this material remains in storage but it is likely that most will be progressively introduced to the sediment flux over the proceeding decades.

As well as directly contributing sediment to the total catchment flux, deep-seated landslides have an underlying control on shallow landslide sediment production. While many shallow landslides occur on gully incision sides, essentially independent of deep-seated landslides, many also occur on deep-seated landslide scarps and these can be considered to be directly controlled by the landslide complex with which they are associated. Rainfall triggered shallow landslides receive much attention as an erosion process (e.g. Brooks et al., 2002; Hennrich and Crozier, 2004) and the widespread occurrence of these in steep hill-country areas has been significantly exacerbated by anthropogenic landscape modification in the form of large scale forest clearance

(Wilmshurst, 1997). In the area of the Hawke's Bay study site there is very little native forest cover remaining and this is inferred to have an influence on the activity of both shallow and deep-seated slope failure. The activity of deep-seated slope failure is likely to have increased to some degree following the removal of the forest canopy due to the increased amount of precipitation reaching the ground surface and becoming available to secondary permeability flow paths, and possibly the decreased soil cover leading to enhanced rates of rock material degradation.

It is clear in these catchments that there is a contribution and underlying control by deep-seated slope failures on sediment production that should not be ignored in any consideration of the sediment flux mechanisms in this setting.

6.1.4 Consideration of a landscape evolution model

The evolution of catchments on the southeastern margin of the Maraetotara Plateau could be numerically modelled to consider how they develop both spatially and temporally. Crucially this would allow consideration of the sensitivity of the system to variations in rates of short and long-term tectonic and climatic forcing. Factors such as uplift rate, earthquake recurrence, critical material strengths and incision rates could be varied and the rate and final form of the landscape would then indicate the importance of these variables and the accuracy of measured or assumed values.

A numerical landscape evolution model would need to have processes and materials described in quantitative terms. Table 6.1 summarises parameters considered to be necessary for a preliminary quantitative landscape evolution model and the initial values which could be applied to them. These values may not all be directly applicable to the Makara Basin setting, however, they can be considered as a starting point from which the sensitivity of the system to specific parameter variation can be assessed.

Parameter	Term	Value	Source	Notes
Rock mass defects	Joint set orientation	Dip 50°, 30° either side of the fold axis	Defect survey (this study)	Dip of defect planes may vary ±15°
	Joint set spacing	45 m	Observation	Difficult to quantify May vary ±10 m
	Joint tensile strength	0 kPa	Field observation	qualitative
	Critical Stratigraphic Horizon spacing	70 m	Field observations	Average value varies ±30 m
	Critical Stratigraphic Horizon strength	$\Theta'_R = 4°$ $C'_R = 10$ kPa	Laboratory testing	Varies: Θ'_R ±1°, C'_R ±5 kPa
Structure	Syncline axis orientation	Aligned to 055°	Figure 2.12	Assume no plunge
	Folding	6 km wide fold	Figure 2.12	Assume symmetric fold
Uplift Rate	Rate	2 m/kyr	(Lewis, 1980; Pillans, 1986; Hull, 1987)	Varies: ±1 m/kyr
Incision Rate	Rate	2 m/kyr	Assumed equivalent to uplift rate	Varies: ±1 m/kyr Note assumption may be invalid due to influence of climatic factors
Seismicity	Average strong ground motion	0.35g	(Stirling et al., 2002)	Peak ground acceleration
	Recurrence interval	150 years	(Stirling et al., 2002)	

Table 6.1: Parameters considered necessary for the development of a quantitative landscape evolution model. A range for values has been provided where possible.

Assumptions needed for numerical modelling include:

- The peak strength of the intact Makara Formation alternating sandstones and mudstones is not relevant to landscape evolution if a rate of stream incision can be assumed. All strength related to slope failure is dependent on rock mass strength

- Measured conjugate joint sets retain their orientation and spacing throughout the lateral extent of the sequence

- Long-term tectonic forcing is represented by a constant uplift rate punctuated by periods of accelerated landscape uplift

- Deep-seated landslides only initiate under the influence of earthquake generated strong ground motion, when slide masses are kinematically admissible

- A slide mass is kinematically admissible when a critical stratigraphic horizon is exposed in a slope

- Full failure of the landslide occurs coincident with the earthquake (co-seismic)

- Once a landslide fails the debris is all removed by the fluvial system

- Landslides will activate with each episode of strong ground motion, and will persist to be active until the failure surface is fully exhumed or is passed through the ridge top

- The influence of the Te Aute limestone on the progression of catchment head incision would be replicated by having a layer capping the model which has an equivalent rock mass strength to the Makara Formation rock mass, but which is defined by lower rates of stream incision into bedrock; and

- Northeast trending normal faults with minor vertical offsets have a negligible influence on landslide development.

While this study has focused on defining and quantifying the internal detail and dimensions of the rock mass which will define a landscape evolution model, the development of the model will require specific boundary conditions. This aspect has not been considered in detail, however, a suggested platform could be a ~10 km x 10 km x 1 km rock mass, either being uplifted and deformed by thrust structures to the west and east, or simply bounded by normal faulting to the east. These are details that need to be addressed by a person versed in the practicalities of writing this type of numerical landscape evolution model.

While the time constraints for this Master of Science study precludes development of such a numerical model, it is hoped that with the data collected and the analysis presented here, it will now be possible to proceed to the next phase of research, involving the development of an approach to numerical modelling, as this will greatly contribute to the understanding of landscape sensitivity to rates of tectonic and climatic forcing over various time scales.

6.2 Discussion on the influence of the Ella Landslide on landscape development in Kate Valley

The geologic and geomorphic setting of the North Canterbury field site compares well with the Hawke's Bay field site. Both coastal landscapes are evolving deeply incised catchments in weak rock material and are influenced by deep-seated slope failure. However, the extent of deep-seated slope failure in the Hawke's Bay field site is far greater than in North Canterbury. In the Hawke's Bay field site critical stratigraphic horizons occur at regular (somewhat predictable) intervals through the Makara Formation, and this is the primary control on the significant extent of deep-seated slope failure. In the soft rock sequences which underlie the north Canterbury field site there are inferred to be only occasional critical stratigraphic horizons, which limits the occurrence of deep-seated landslides.

Uplift of up to 2.16 mm/yr (Nicol et al., 1994) is reflected in marine terraces, which are in places also tilted, stepping up and away from the coast, and also by the deeply incised stream networks. Prior to the occurrence of the Ella Landslide, the Kate Stream incision would likely have extended significantly further up the catchment probably at least to Cass Landslide (refer Figure 3.1), and the geomorphic development of slopes in the catchment above Ella Landslide would have been significantly different as slope erosion processes were previously directly coupled with the fluvial system. The Ella Landslide caused a barrier to sediment removal from the upper catchment and caused approximately 60 m thickness of sediment aggradation (Geotech Consulting Ltd, 2002) that forms the alluvial plain present today. The presence of this alluvial plain means that where previously there was a debris removal mechanism active at the toe of hillslopes in the form of the fluvial system (stream coupled to hillslopes), lower slopes are now supported by the valley fill which has a significant effect on the style and activity of hillslope processes.

When a stream is actively incising, and defines a deeply incised gully, such as that below the Ella Landslide, hillslope erosion processes are active from stream level to ridge top and hillslope denudation has been directly linked to stream incision rates in some landscapes (e.g. Burbank and Anderson, 2001). Now that the slope toe in the upper catchment is supported by valley fill, erosion occurs primarily as shallow landslides to the depth of weathering and soil creep. As Kate Stream eventually incises through and past the Ella Landslide debris the hillslopes will once more become coupled to the fluvial system and rejuvenate the denudation process.

The overall effect of the Ella Landslide event on the geomorphic development of the Kate Stream catchment can be considered from several aspects:

- Hillslope denudation processes in the middle catchment are significantly retarded by effective decoupling of slopes from the fluvial system,
- The sediment flux from Kate Stream to the offshore basin system on a $10^2 - 10^3$ year scale has been significantly depleted as the majority of sediment available in the upper catchment is currently kept in storage; and
- On a $10^4 - 10^5$ year scale the sediment production flux from the catchment may be increased due to the significant input from such large slope failures.

It is clear that the occurrence of isolated slope failures has a significant impact on the development of the Kate Valley catchment which contains two significant size slope failures, the Ella and Cass landslides. Although the impact of the Cass Landslide on catchment development has not been assessed, the relationship of this to the timing of Ella Landslide and to catchment development can be inferred. The Cass Landslide is likely to be several thousand years older than Ella Landslide based on comparative degradation of the landslide scarps and it is possible that these two landslides are related to different episodes of incision driven by rapid base level lowering. The Cass landslide would have been able to initiate once incision by Kate Stream had exposed a similar weak stratigraphic horizon to that defining the failure surface of the Ella Landslide. The failure of Cass Landslide would have affected the catchment above it in a similar way to how the Ella Landslide has affected the upper Kate Stream catchment by damming the stream and developing some form of alluvial plain (this is inferred from vertical air photo interpretation, NZ Aerial mapping run 1824/48-51). Once incision had progressed through the Cass Landslide debris it removed most of the valley fill material buttressing lower slopes upstream of the slope failure and the slope once again became stream-coupled. At some point after the removal of the majority of valley fill, the Ella Landslide event occurred (possibly following a period of accelerated base level lowering), and sediment removal was further retarded. Incision has not yet progressed through the Ella Landslide debris, however, eventually the significant volume of material stored in the mid – upper catchment will be eroded out of the stream system. It is possible that a further period of accelerated base level lowering in the future will expose further critical stratigraphic horizons and allow more deep-seated slope failure to occur.

The paucity of critical stratigraphic horizons within the Tokama Siltstone in comparison to the Makara Formation is controlling the limited spatial occurrence of deep-seated slope failures in the North Canterbury field site. Despite this limited occurrence, these failures have a significant impact on catchment evolution and partially define the rate and style of catchment development.

6.3 Implications for mountain range scale landscape evolution models

The development of mountain range scale numerical landscape evolution models which realistically consider the interaction of tectonic and geomorphic processes is an interdisciplinary goal and in the past models have commonly focused on one process at the expense of simplifying the other, often to the consternation of respective disciplines (e.g. Merrits and Ellis, 1994). As a generalisation, geomorphologists may focus on portraying the geomorphic processes as accurately as possible while simplifying long-term tectonic forcing to spatially uniform steady uplift rates or alternating high and low uplift rates, while geophysicists may be inclined to consider complex tectonic process and isostatic response but simplify geomorphic processes to an instantaneous lowering of unstable hillslopes above a defined threshold.

Some levels of complication within a model may be precluded by the specific setting, for example Anderson (1994) included the possibility of crustal flexural response to topographic unloading in a two dimensional model to describe the evolution of the Santa Cruz mountain range, but subsequently showed that due to the scale of the mountain range no such response was active. The inclusion of some specific processes is, however, crucial if the development of a topographic range is to be replicated to any extent by realistic mechanisms. To model the evolution of the Southern Alps Koons (1989) included channel incision and diffusion to encompass all geomorphic action and found that the required rate of diffusion to replicate slope lowering was significantly higher than realistic because landsliding and other geomorphic processes were all represented by diffusion. The specific processes considered relevant to a given model are intrinsically scale dependant. For a long-term orogenic scale model the effect of upper crustal faulting may be considered irrelevant (Densmore et al., 1998), while at the individual hillslope scale the effect of isostatic rebound due to material removal will certainly not apply. It is what may be defined as the mountain range scale, (e.g. Burbank and Anderson, 2001), that is of interest to this project, where all processes are accounted for, and which affect landscape development at the scale of the major catchments systems.

The three dimensional model ZSCAPE (Densmore et al., 1998; Ellis et al., 1999) probably goes as far as any mountain range scale landscape evolution model to date by including realistic tectonic and geomorphic processes. ZSCAPE considers the influence of upper crustal faulting and assumes that deformation is the result of repeated rupture of upper crustal faults. In an application to the Basin and Range province in the Western United States for a 10 km by 10 km area with 100 m grid spacing Densmore et al. (1998) consider a $M_W = 7.0$ earthquake every 500 years on normal faults producing surface displacements between 140 cm of subsidence to 30 cm of uplift with three dimensional accumulated tectonic deformation. Erosion and deposition in the model are dictated by conservation of mass related to the rate of change of the surface elevation with respect to spatial gradients in sediment flux. Figure 6.8 shows two images from the ZSCAPE programme considering the evolution of a normal-fault-bounded mountain taken 500 kyrs apart.

The geomorphic processes which allow this are:

- Regolith production, for which the rate increases up to a defined regolith thickness and then decreases exponentially

- Regolith transport, (including creep, slope wash, rain splash and animal activity) which is defined as a slope dependant linear diffusion process. Advective transport processes such as shallow landsliding are modelled by allowing the diffusion rate to increase exponentially as hillslopes approach a critical angles

- Fluvial processes, where alluvial and bedrock channel behaviour is distinguished based on material availability and the fluvial sediment transport flux is proportional to available stream power with respect to bed area; and

- Bedrock landsliding, where the rock mass is treated as homogeneous and landslide initiation is controlled by a threshold hillslope height.

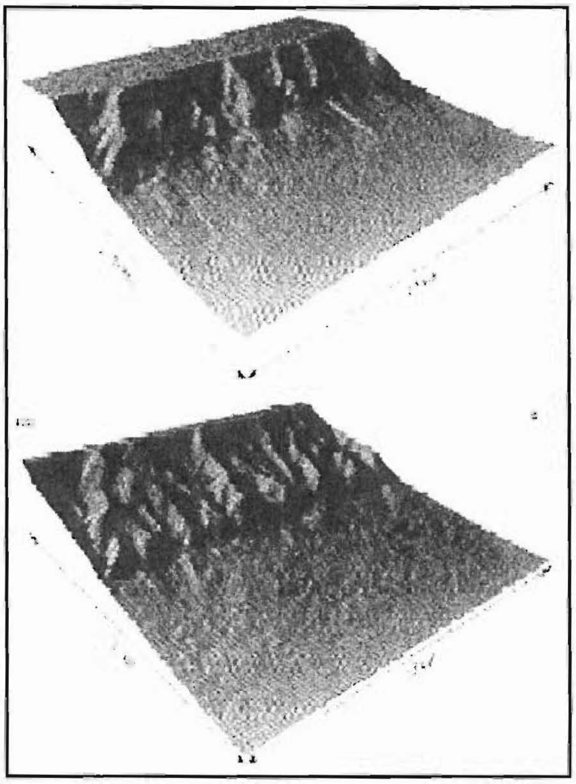

Figure 6.8: Stages of mountain range scale landscape evolution as developed by the landscape evolution model ZSCAPE. Plots are 10km x 10km and shown with 2x vertical exaggeration. The top image is after 500 kyr of landscape development and the bottom image after 1000 kyr. Modified from Densmore et al. (1998).

The model replicates the form of the landscape, being consistent with laboratory and field observations and indicates that bedrock landsliding is an important or even dominant process in the development of this landscape. The medium in which processes act to replicate landscape form in ZSCAPE is treated as homogeneous, however, the lack of any representation of rock mass defects which may control slope stability and morphology is acknowledged (Densmore et al., 1998). The control of rock mass defects on geomorphic development is reasonably well documented (e.g. Selby, 1980; Weissel and Seidl, 1997) and will affect the replication of smaller scale geomorphic features such as slope angles, aspects and the location of stream beds.

Both the landscapes considered in this study, and by implication the larger population of landscapes developing in Tertiary soft rock terrain in New Zealand, are strongly influenced by the occurrence rock mass defects, including joint and fault sets and bedding parallel defects. It is unlikely that the catchments in the Hawke's Bay study area could be replicated to any extent by modelling their development in a homogeneous medium. To numerically model any landscape in bedded Tertiary soft rock sequences, and realistically replicate the hillslope geometry and deep-seated landslide distribution, a heterogeneous and discontinuous medium is considered to be essential. This would need to define the three-dimensional components of defect sets highly oblique to bedding and bedding parallel weaknesses.

6.4 Chapter summary

Deep-seated or bedrock landsliding is a dominant factor in hillslope denudation and landscape evolution for a variety of landscapes including those developing in Tertiary soft rock terrain. Compared to the many mountainous landscapes in which this mass movement process has been considered to be having a significant impact on landscape development, topography in Tertiary soft rock terrain is characterised by low to moderate relief hill country. The influence of the Ella Landslide on the geomorphic development of the Kate Stream catchment in North Canterbury highlights the long-term effect that even isolated instances of deep-seated slope failure can have on catchment evolution. Existing numerical models of landscape evolution in mountainous terrain generally consider bedrock landslide initiation to be controlled by base level lowering and a defined slope height threshold. It is recognised in this study that rock mass defects have a controlling influence on many aspects of landscape development in Tertiary soft rock terrain including deep-seated landslide geometry and spatial distribution, slope geometries and the location and development of fluvial networks. While existing landscape evolution models that include deep-seated (bedrock) landslides as a mass wasting process consider their occurrence to be stochastic, the spatial control on the initiation and maintenance of deep-seated landslide complexes by inherent rock mass defects and the temporal control by earthquake triggering means that they are implicitly a non-stochastic or spatially and temporally predictable process.

To consider the evolution of the catchments at the southeastern margin of the Maraetotara Plateau in terms of the influence of deep-seated landslides, this study has quantified parameters such as critical rock mass properties (e.g. strength, spacing and orientation of defects including both conjugate joint and fault sets and critical stratigraphic horizons) and the level of strong ground motion required for landslide triggering. Other parameters required can be defined from published data including; uplift rates, structural characteristics peak ground acceleration magnitude and recurrence, while parameters such as stream incision rates into bedrock must be inferred.

The aim of providing parameters for the development of a numerical landscape evolution model is that the development of such a model would allow for the sensitivity of the landscape to tectonic and climatic forcing factors (e.g. uplift rates, earthquake magnitude and recurrence) and parameters that are inherently difficult to quantify such as stream incision rates to be tested.

Chapter Seven

7.0 Summary and Conclusions

7.1 Introduction

Two study areas have been chosen, in North Canterbury and southern Hawke's Bay, which are representative of a much larger population of small to medium sized catchments throughout the large areas of New Zealand underlain by Tertiary soft rock terrain. By assessing the role of deep-seated landslides in the evolution of these representative study catchments, inferences can be drawn about the geomorphic and morphologic development of a significant component of the New Zealand landscape, in which the widespread occurrence of deep-seated failures is well documented. To assess the role of deep-seated landslides in the evolution of these landscapes with the aim of providing inputs for the development of numerical landscape evolution modelling, it is necessary to understand and quantify the specific controls on the geometry, failure mode and the spatial and temporal occurrence and persistence of the deep-seated slope failures within them.

The deeply incised, small to moderate sized (15 – 20 km^2) stream catchments at the southeastern margin of the Maraetotara Plateau, southern Hawke's Bay, provide an excellent example of a situation where landscape evolution is controlled by numerous deep-seated slope failures. This landscape is developing in the strata of the Miocene age Makara slope basin, uplifted and emergent on the highest accretionary ridge of the Hikurangi Margin frontal wedge. The basin fill sequence consists of a thick succession (~1000 m) of alternating sandstone-siltstone/mudstone flysch (Makara Formation), unconformably capped by ~10 m of erosion resistant limestone (Te Aute Formation). During Miocene deposition of terrestrially derived sediment, the Makara Basin also received periodic pulses of volcanic ash that were incorporated into the succession as discrete and laterally continuous horizons. Strata are folded into a thrust fault bounded syncline, trending NE-SW, and are also affected by SE dipping normal faults striking NE-SW, with NW dipping conjugate faults. The main study area is focused around the core of this syncline where dips of between 0° and 15° control the failure mode of the numerous examples of deep-seated landslides failing as low to moderate angle retrogressive block slide complexes and wedge failures.

In North Canterbury the small (~4 km^2) Kate Stream catchment provides an example of a catchment where the occurrence of occasional deep-seated slope failure has a significant impact on the development of the catchment system. This landscape is forming in a Tertiary succession uplifted and folded into thrust fault driven, asymmetric anticline/syncline pairs that have been subject to periods of accelerated base level lowering, represented by a deeply incised stream network and uplifted (and in places tilted) marine platforms. The catchment contains a large, deep-seated landslide (Ella Landslide) that has a basal failure plane defined by a 5 – 10 mm thick, Kaolinite rich, pre-sheared stratigraphic horizon.

7.2 Origins and implications of critical stratigraphic horizons within the landscape

The deep-seated landslides considered in this study have basal failure surfaces defined by bedding parallel horizons which have been defined as *critical stratigraphic horizons*. These discrete and thin horizons have a high strength contrast with the enclosing dominant lithology, are pre-sheared, and are laterally continuous throughout the succession. The origin of critical stratigraphic horizons in soft rock successions is proposed to be dominantly controlled by lithological variation. In this study the observed critical stratigraphic horizons in the respective study areas have different origins, and other documented cases of stratigraphically controlled failure surfaces for deep-seated landslides provide further sedimentary origins for critical stratigraphic horizons.

In the Hawke's Bay study area the critical stratigraphic horizons are defined by tuffaceous beds deposited during sediment accumulation in an offshore slope basin. During burial, compaction and subsequent exhumation, these tuffaceous beds did not "overconsolidate", as the surrounding Makara Formation sandstone and siltstone/mudstone units did, but remained as an engineering soil. By the time the succession was fully emergent the tuff layers defined a distinct strength contrast, as weak horizons within the stratigraphic sequence. Geotechnical testing of the critical stratigraphic horizon material defines residual strength parameters of effective cohesion (C'_R) = 3.8 – 14.2 kPa and effective friction angle (Θ'_R) = 2 – 5°, by comparison the peak strength of the adjacent siltstone/mudstone is inferred to be in the order of Θ' = 35°. The material of the critical stratigraphic horizons is considered to be at residual strength as the tuffaceous beds have experienced some shear displacement in the accommodation of rock mass deformation. What defines specific tuffaceous horizons as critical stratigraphic horizons in the succession is the shear displacement that has occurred within them, and this is proposed to be due to a combination of: i) flexural (layer on layer) slip during folding, which utilised the strength contrast of the weak tuffaceous material to accommodate folding by bedding-parallel shear deformation within these horizons; and, ii) rock mass adjustment due to thrust fault propagation, where displacement on northwest dipping thrust faults induces localised stress concentrations that are accommodated by adjustment of the rock mass on pre-existing defect surfaces (bedding planes, joints and faults). It is the shearing of specific stratigraphic horizons which defines them as "critical" to the occurrence of deep-seated slope failure, and while the tuffaceous horizons are somewhat regularly spaced by depositional nature, distributed strain in the rock mass may also contribute to the generally regular spacing of these through the stratigraphic column as specific bedding horizons shear to accommodate, and distribute, rock mass deformation. It is primarily the occurrence and nature of these critical stratigraphic horizons which has lead to the numerous instances of deep-seated slope failure throughout this study area.

In the Kate Stream catchment, North Canterbury, there are very few occurrences of deep-seated slope failure and this reflects the limited occurrence of critical stratigraphic horizons within this succession. The horizon exposed in Kate Stream has been characterised by laboratory testing which shows: i) the 5 – 10 mm thick horizon occurs within a ~300 mm thick package of clay rich material and defines a locus of fine grained material within this; ii) the layer contains approximately 15% of the clay mineral Kaolinite, three times the Kaolinite content of immediately adjacent material (enclosing stratigraphic units); and, iii) the microstructure of the intact horizon material confirms what was observed in the field, that the layer has experienced shear displacement. The occurrence of higher concentrations of Kaolinite, and the finer grained nature of this material suggests that it has a syn-depositional origin and is inferred to have been deposited from a "cloud" of fine grained material settling out of the water column and blanketing the sea floor in a low energy, outer shelf environment. The shear displacement on the layer is proposed to relate to flexural (layer on layer) slip during folding. Shear within the material has reduced it to being at or near its residual strength ($C'_R = 2.6 - 2.7$ kPa, and $\Theta'_R = 16 - 21°$) which is a significant contrast with the peak strength of the dominant lithology ($C' = 176$ kPa, and $\Theta' = 37°$).

Basal failure surfaces for deep-seated landslides that might be considered as critical stratigraphic horizons are well documented in soft rock successions throughout New Zealand and the world (e.g. Skempton, 1964, 1966; Bjerrum, 1967; Sugden et al., 1977; Pinckney et al., 1979; Coombs and Norris, 1981; Barton, 1984; Pettinga, 1987a; Barton, 1988; Bell and Pettinga, 1988; Fell et al., 1988; Pettinga and Bell, 1992; Prebble, 1992; Hutchinson and Anonymous, 1995; Hart, 2000; Hamel and Hart, 2001). Sedimentary origins for critical stratigraphic horizons may include: i) a permeability contrast; ii) an increase in clay content; iii) a change in clay mineralogy; or, iv) a strength contrast such as increased cementation or lithological bedding. All these sedimentary characteristics can provide a shear strength contrast or "defect" in the stratigraphy which may cause a particular horizon to shear in preference others. In addition, a stratigraphic defect that has no clearly defined material contrast either side (i.e. the sedimentary characteristics appear to be the same above and below) may be considered as a "bedding parting" which may simply be defined by a brief pause in sedimentation.

Implications for landscape evolution

In the Kate Stream catchment the occurrence of a weak clay-rich critical stratigraphic horizon, in an otherwise moderately strong silty-fine sand dominated succession, has allowed deep-seated slope failure to occur and this has significantly affected the evolution of this catchment. The Holocene Ella Landslide has formed a landslide dam and decoupled hill slopes in the middle catchment from the fluvial system. It is most likely that without the occurrence of this horizon in the stratigraphy there would be no deep-seated slope instability, and development and morphology of the middle catchment would be significantly different.

In the catchments adjacent to the Maraetotara Plateau the occurrence of very weak (Θ'_R = 2 – 5°), regularly spaced (~70 m) and laterally continuous critical stratigraphic horizons has a controlling influence on the spatial distribution of deep-seated landslides, rates of hillslope lowering and the morphology of the landscape. Deep-seated landslide geometry is controlled by the combination of the basal failure surface defined by a critical stratigraphic horizon and the headward release surfaces defined by conjugate defect sets. Planimetrically extensive areas of the mid-upper portions of the catchments in this study area are affected by deep-seated landslide complexes which can be correlated between catchments to be failing on the same stratigraphic horizon. The regular stratigraphic spacing of critical stratigraphic horizons, combined with their lateral extent and significant control on deep-seated slope failure defines a stepped slope profile that characterises the landscape morphology in these catchments. Without the presence of critical stratigraphic horizons in the Makara Formation the mass wasting processes that are controlling catchment and slope evolution and morphology would be significantly different.

It is proposed that the critical control on the widespread occurrence of deep-seated slope failure in other areas of New Zealand formed in Tertiary soft rock terrain is the presence of critical stratigraphic horizons and that without these, large deep-seated slope failures would be unable to occur. Critical stratigraphic horizons can then be considered one of the critical influences on the development and morphology of landscapes in Tertiary soft rock terrain.

Implications for engineering practice in soft rock terrain

Stability modelling of two deep-seated soft rock landslides provides an indication of the sensitivity of slopes in bedded Tertiary soft rock sequences to factors such as material strength, failure plane angle, release surfaces, hydrological conditions and seismic ground motion. Under realistic hydrological conditions (slope failures involving ridges are unlikely to have saturated slide masses) it appears that neither of these slopes would fail in a static situation (factors of safety greater than 1.3). The occurrence of perched water tables having an influence on slope destabilisation has been discounted in both field areas based on field mapping of lithological characteristics. In both situations it is the very high strength contrast between the dominant lithology and the thin weak layers, or critical stratigraphic horizons, which not only defines the location of these landslides but has a controlling influence on slope stability.

The occurrence and criticality of bedding parallel failure surfaces to bedrock slope stability is well documented in engineering practice (e.g. Stout, 1977; Fell et al., 1988; Yue and Lee, 2002). The extensive lateral continuity and the overriding influence on landslide location and initiation of critical stratigraphic horizons within the Makara Formation, may have significant implications for large engineering projects in any Tertiary soft rock terrain where critical stratigraphic horizons are present. If an engineering project covers a large area (e.g. a roading corridor or hydroelectric scheme) and is to be undertaken in soft rock terrain, the implications from this study could be applied at all stages from project feasibility to construction. A deep-seated landslide (or several) can cause significant time and cost overruns for an engineering project (e.g. the Clyde dam project, Gillon and Hancox, 1992) or lead to the complete abandonment of the project (e.g. the Vajont Dam failure, Hendron and Patton, 1987). The implications of this study for engineering practice are that the use of detailed topographic and stratigraphic information combined with local seismic hazard data can be used to predict where critical stratigraphic horizons occur in the landscape and subsequently assess the sensitivity of slopes to deep-seated failure.

Critical horizons can be recognised in the stratigraphy by undertaking detailed face logging, core drilling and/or locating the failure surface of pre-existing bedrock landslides. As these layers are typically only a few to several millimetres thick and their location is critical, projecting them through the landscape requires very detailed topographic information (a method such as aerial laser altimetry, or LIDAR, would provide this). Combined with information on joint sets, critical material strengths and groundwater conditions it would be possible to develop very accurate slope models to consider static and dynamic slope stability.

This detailed consideration of slope stability would allow either avoidance of particularly critical sites or allow remediation work to be undertaken. The cost of undertaking such an investigation would not be insignificant, however, in the kind of terrain (highly dissected soft rock landscapes) where these landslides occur the recognition of potential deep-seated slope failures could have major implications for project viability, and the expenditure at the site investigation stage would be well justified.

7.3 Development of parameters for a numerical landscape evolution model

A numerical landscape evolution model that broadly replicates the development of the landscape at the southeastern margin of the Maraetotara Plateau and enables the assessment of the sensitivity of this landscape to varying rates and values of specific parameters, needs to consider the influence of tectonic and geomorphic processes on a heterogeneous rock mass.

Based on the geometry and stability of a block failure that is representative of deep-seated slope failure in the study area, and field mapping of conjugate defect sets and the spacing and orientation of critical stratigraphic horizons the spatial extent of rock mass discontinuities can be defined. Critical stratigraphic horizons are typically spaced at 50 – 80 m, with an average of 70 m. While average joint spacing may be in the order of a few metres, based on the observed planimetric size of discrete slope failures the spacing of "critical" defects which are likely to define slide block release surfaces is in the order of 40 – 50 m, and these typically dip at ~50°. This means that the landscape can be divided into a discontinuous medium defined by ~2 x 10^5 m^3 blocks. The only defining strength with respect to slope stability in an unsupported rock mass is considered to be the low shear strength of the critical stratigraphic horizons, as joints have no tensile strength.

Other significant controls on landscape evolution include tectonic and climatic forcing. Stream networks have become deeply entrenched due to base level lowering related to both the sustained tectonic uplift and orbitally forced glacial/ interglacial controlled sea-level variation (long-term climatic forcing). Since the Makara Basin became emergent (Pliocene to early Pleistocene) it is long-term tectonic and climatic forcing (average landscape uplift rate of 2 mm/yr) which are the underlying drivers of sub-aerial landscape development processes, as stream incision progressively exposes critical stratigraphic horizons and allows slopes to become unsupported. Short-term landscape forcing is considered in terms of the triggers for slope degradation processes, specifically deep-seated landslides. While short-term tectonic forcing refers to landscape perturbation in the form of large magnitude earthquakes (which may be defined from published seismic hazard data, e.g. Stirling et al., 2002), short-term climatic forcing refers to annual to decadal climatic variability. Numerical slope stability modelling indicates that earthquake generated strong ground motion is required to trigger deep-seated slope failure. While the full failure of a slide mass may occur over periods of decades to centuries, coincident with adverse hydrological conditions, in geomorphic time frames such slope failures are essentially co-seismic.

As the location of the basal failure surface of deep-seated landslides is controlled by the occurrence of critical stratigraphic horizons, deep-seated slope failure will not occur until these are exposed in the landscape. The stratigraphic succession is exposed as stream networks incise into bedrock and critical stratigraphic horizons will eventually be exposed at stream base level. It is at this point that deep-seated landslide complexes are able to initiate, controlled by strong ground motion during large magnitude earthquake events. Once initiated landslide complexes will most likely persist in the landscape, with ongoing activity controlled by subsequent earthquake events. As stream incision rates increase, controlled by tectonically and climatically forced acceleration of base level lowering, landslide complexes will be come perched within the catchment but will continue to be active for thousands to tens of thousands of years. As lower critical stratigraphic horizons are exposed by renewed stream incision, and landsliding is activated on these, upper critical stratigraphic horizons will persist to be active until they are either fully exhumed, or they are passed out through the (lowering) ridge top. It is the continued initiation and persistence of deep-seated slope failures, controlled by rock mass defects and tectonic and climatic forcing, that has a dominant control on the evolution and morphology of these catchments as they develop within this specific succession.

By quantifying controls on deep-seated landslide initiation, geometry and behaviour in the specific situation of the catchments developing in the Makara Basin sequence, it is possible that the initiation and persistence of these in time and space can be considered as non-stochastic. This will be important in a numerical model of landscape evolution where the realistic scenario of rock mass defect controlled deep-seated (bedrock) landslides initiating with the confluence of critical conditions will allow an assessment of the sensitivity of the landscape to specific parameters such as: uplift rate variability; incision rates; earthquake recurrence; earthquake magnitude; and, rock mass strength.

7.4 Major conclusions of this study

- Rock mass defect controlled, deep-seated landslides as a mass wasting process have an underlying and possibly controlling influence on the development of catchments, and subsequently landscapes, in Tertiary soft rock successions in New Zealand.

- The geometry and failure modes of deep-seated landslides in Tertiary soft rock terrain are controlled by rock mass defects, specifically the strength and orientation of critical stratigraphic horizons and conjugate joint/fault sets.

- Critical stratigraphic horizons within the rock mass can define the basal failure surfaces for numerous instances of deep-seated slope failure and so have a significant influence on landscape evolution of catchments developing in stratigraphic successions including critical stratigraphic horizons. Critical stratigraphic horizons are characterised as: i) having a high strength contrast with the surrounding material; ii) having experienced some shear displacement within the intact rock mass; and, iii) being laterally continuous over geographically extensive areas within the stratigraphic succession.

- Long term tectonic and climatic forcing factors control landscape development by defining the tilting and folding of strata, and controlling the magnitude and fluctuation of base level lowering rates which allow the deep incision of fluvial networks, and the exposure of stratigraphic successions including critical stratigraphic horizons.

- The frequency of occurrence of critical stratigraphic horizons in a succession has a significant impact on the form of the landscape developing within that succession. Deep-seated slope failures may only occur in bedded soft rock sequences where critical stratigraphic horizons are present.

- The temporal initiation of deep-seated landslides in Tertiary soft rock terrain is predominantly controlled by the periodic occurrence of large magnitude earthquakes. The strong ground motion induced by these events leads to the dilation of existing rock mass defects and the displacement of slide blocks whose (critical stratigraphic horizon coincident) basal failure plane is exposed at or above stream base level.

- The full failure of displaced blocks will occur some tens of years later due to enhanced (secondary) permeability, following rock mass dilation, and material degradation (slaking). In the context of landscape evolution on a geomorphic timescale ($10^4 - 10^5$ yrs), however, the initial displacement and full failure or slide mass disintegration can be considered as a single (seismically triggered) event.

- A landscape that is developing within a succession containing numerous and regularly spaced critical stratigraphic horizons will attain a specific overall slope geometry, termed a "stepped landscape geometry". The "treads" or steps are defined by the extent of the various basal failure planes of deep-seated landslides and the "risers" of the steps are defined by the vertical spacing between critical stratigraphic horizons. The stepped landscape geometry will be maintained for as long as the landscape is evolving in the particular succession which contains the critical stratigraphic horizons which act as basal failure surfaces for deep-seated landslides.

- The lowering of slopes will be in a form of "steady state", where the inclusion of stratigraphically lower critical stratigraphic horizons as basal failure surfaces in deep-seated landslide complexes is defined by accelerated stream bed lowering following accelerated base level lowering events. Subsequently the evolution of the landscape and the lowering of catchment slopes are directly coupled to the magnitude and fluctuation of base level lowering rate as controlled by long term tectonic and climatic forcing.

7.5 Potential for subsequent research projects

Subsequent to this project, it is envisaged that the following would be worthwhile areas for research that would be of benefit to the understanding of deep-seated landsliding mechanisms and their influence on landscape evolution in Tertiary soft rock terrain:

- Development of the numerical landscape evolution model discussed. With the quantified aspects of the lithological heterogeneity and tectonic and climatic controls this would allow an assessment of the sensitivity of landscape development to varying rates of long and short term tectonic forcing

- Further research into the sedimentary and tectonic controls on critical stratigraphic horizon development. While this study has shown that an initial sedimentary origin defines the location of subsequent (tectonically induced) shearing, a full investigation into the lithological characteristics of these layers in a wide variety of successions, and a detailed structural analysis of the mechanism of shear development would provide a significant contribution to the understanding of deep-seated slope failures in Tertiary soft rock

- Development and application of a method for using high resolution topographic data to project critical stratigraphic horizons through the landscape to allow a quantitative deep-seated slope stability assessment without large quantities of sub-surface data. This is directly applicable to large engineering projects in soft rock terrain where deep-seated landslides can dramatically affect project feasibility

- Temporal quantification of bedrock incision and landslide chronology using relict landscape features and dating deep-seated landslide events. This may not be feasible in the Makara Basin setting due to the lack of preserved head scarp grabens and similar good dating locations, however in a location such as the Waipaoa Catchment in the Raukumara Peninsula or the Central Rangiteiki region this quantification may be possible using the temporal constraint of the numerous volcanic ash layers recorded in the thick soils mantling slopes. This could provide quantitative field evidence for both down-cutting rates, which are typically poorly constrained, and the relationship between down-cutting rates/events and deep-seated landslide initiation. An assessment of whether populations of deep-seated landslides occur in the landscape with, or at some time following, a base level lowering event would allow calibration of (as yet undeveloped) numerical landscape evolution models which may predict such relationships; and

- Quantification of the input from deep-seated landslides in Tertiary soft rock catchments to the sediment flux, both as direct contributors and as an underlying control on other mass wasting processes. The large sediment flux from catchments formed in Tertiary soft rock terrain is commonly attributed to shallow mass movement and the role of deep-seated slope failure is underestimated or overlooked. Confirmation of the contribution of deep-seated landslides to the sediment flux as being significant would support the hypothesis that deep-seated landslides control, and are the driving factor for catchment evolution in this terrain type.

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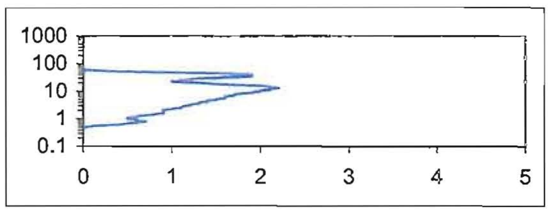
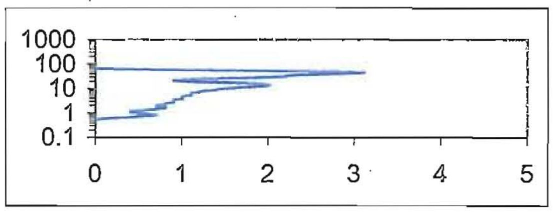
Appendix I

Grainsize analysis curves

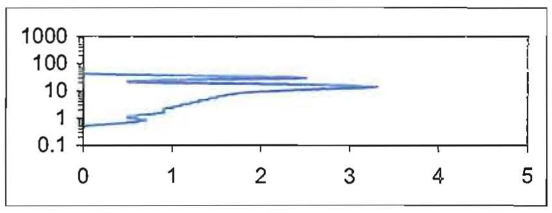
+560 mm



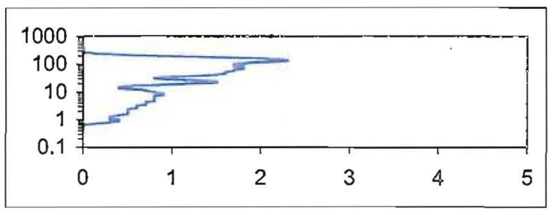
+540 mm



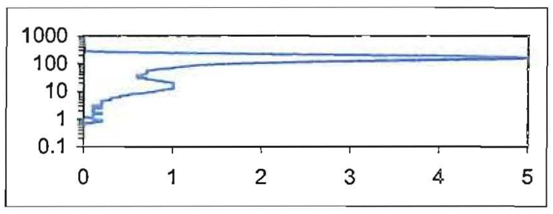
+530 mm



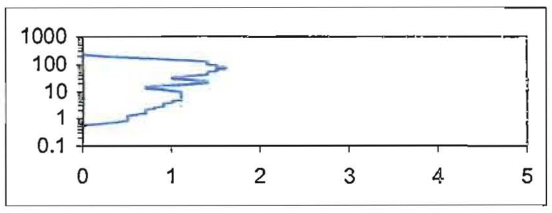
+460 mm



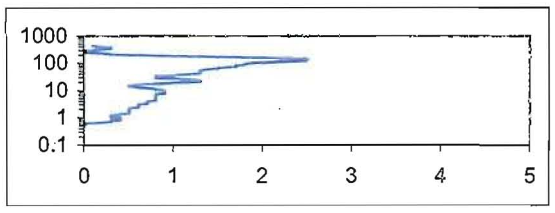
+430 mm



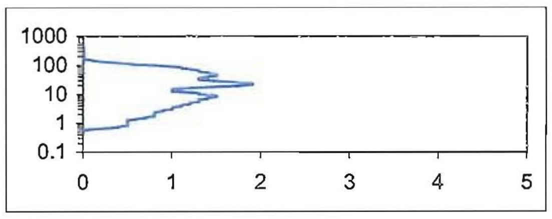
+350 mm



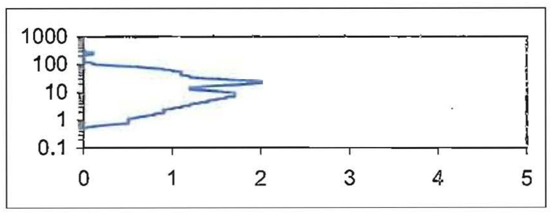
+280 mm



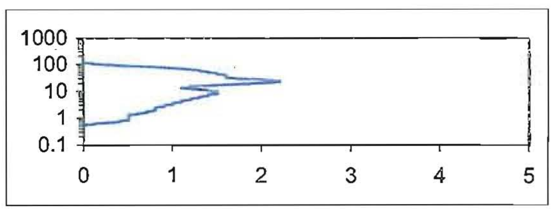
+115 mm



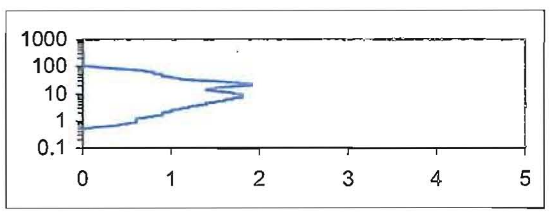
+60 mm



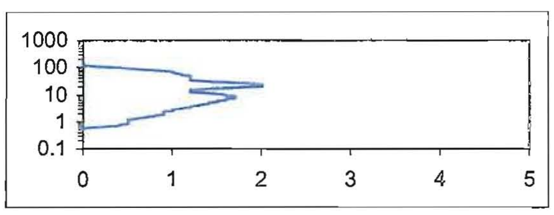
+40 mm



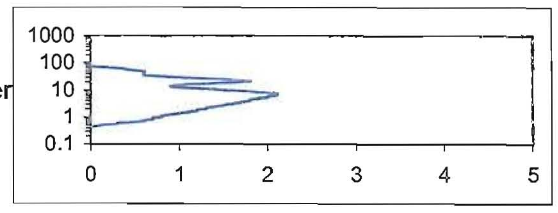
+20 mm



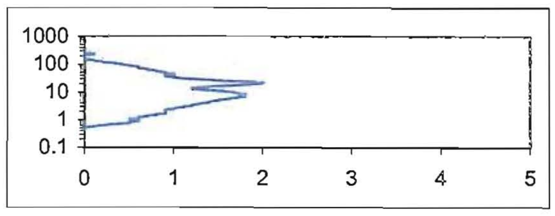
+10 mm



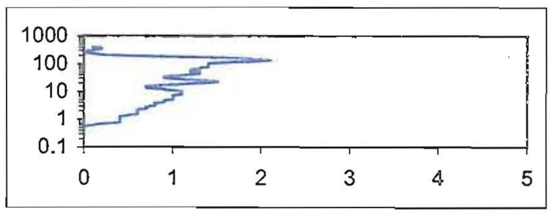
Clay Layer



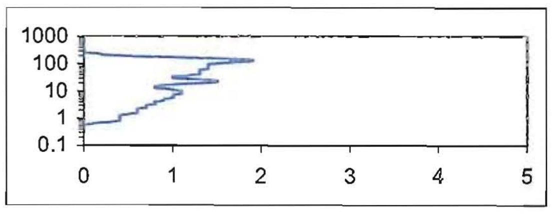
-10 mm



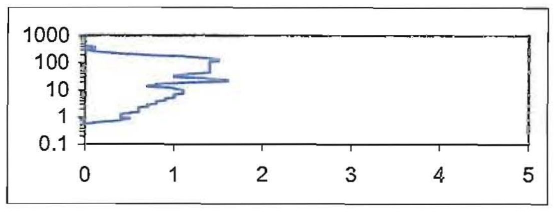
-20 mm



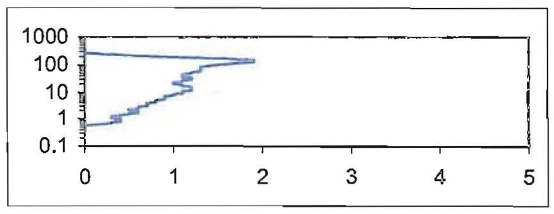
-40 mm



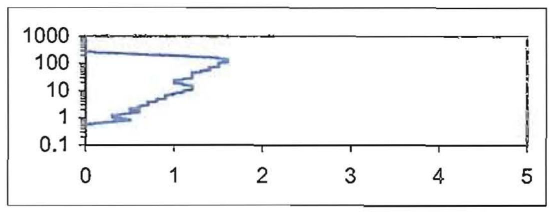
-60 mm



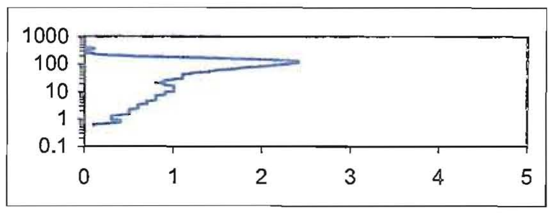
-80 mm



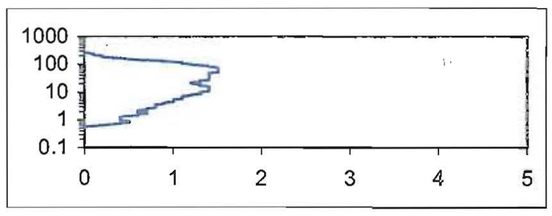
-100 mm



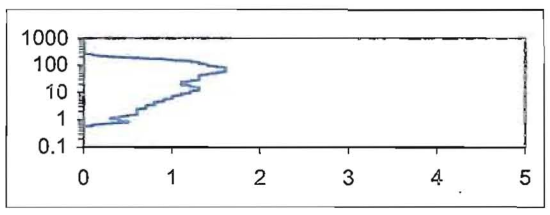
-150 mm



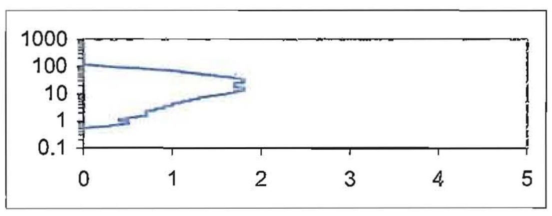
-170 mm



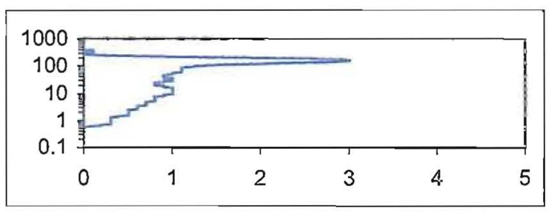
-190 mm



-250 mm



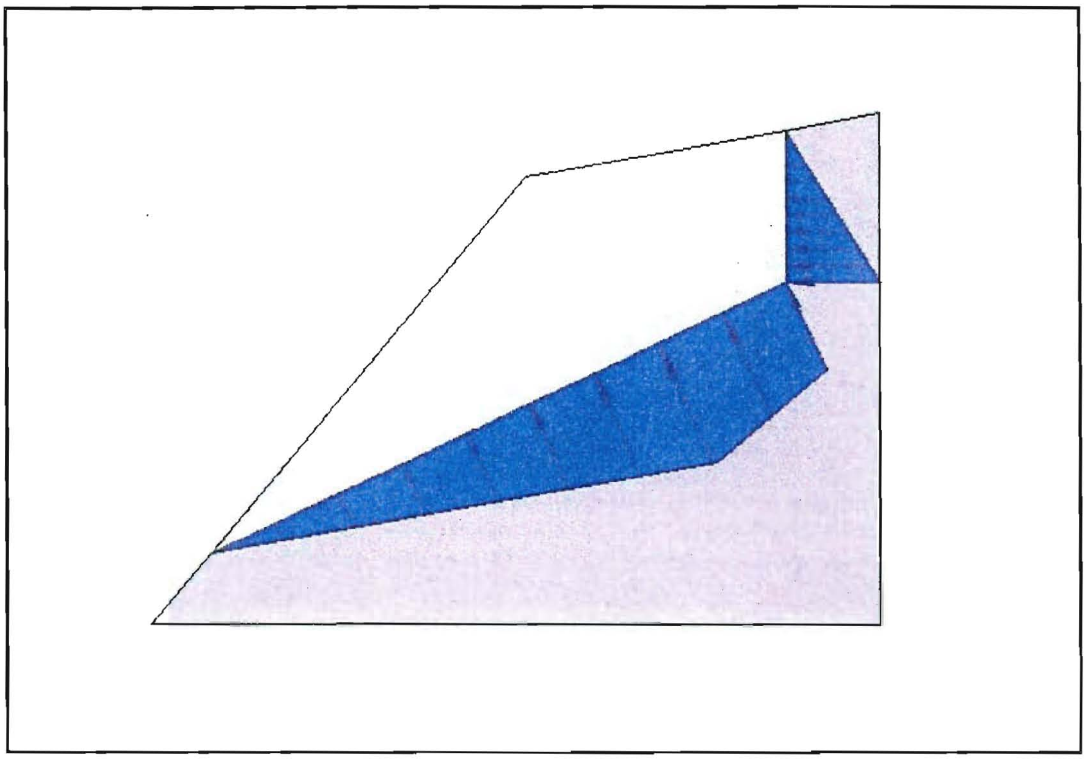
-350 mm



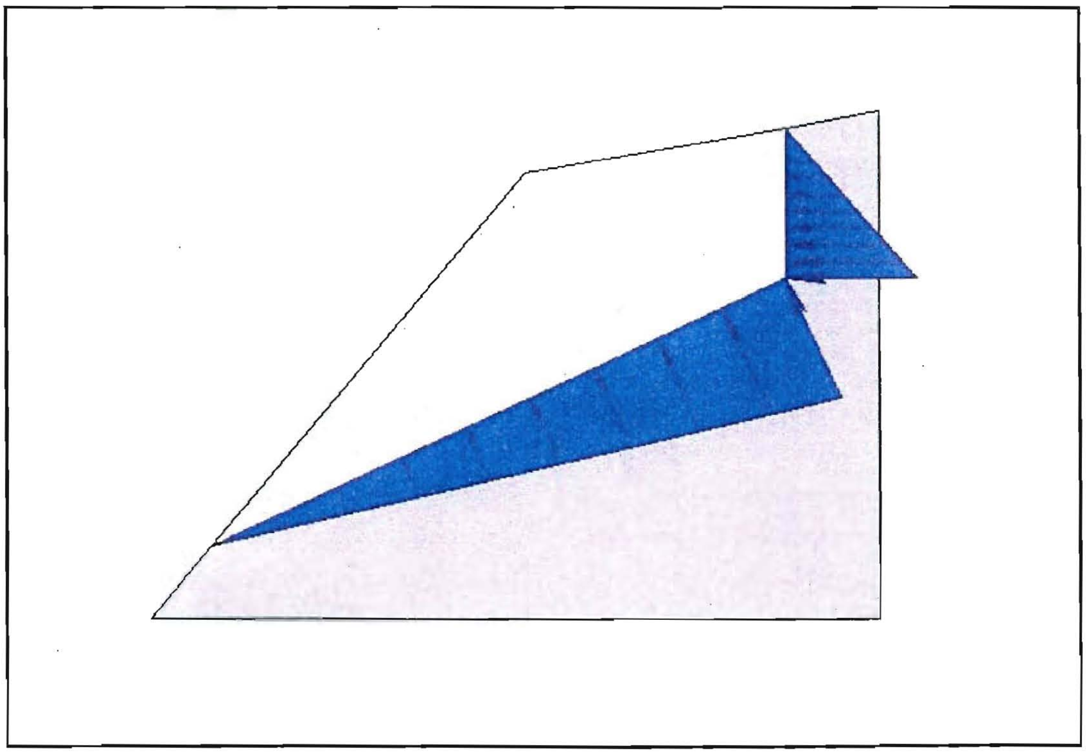
All plots axis are the same. Vertical (Y) axis shows log of grainsize in microns. Horizontal (X) axis shows the volume frequency for each grainsize as %.

Appendix II

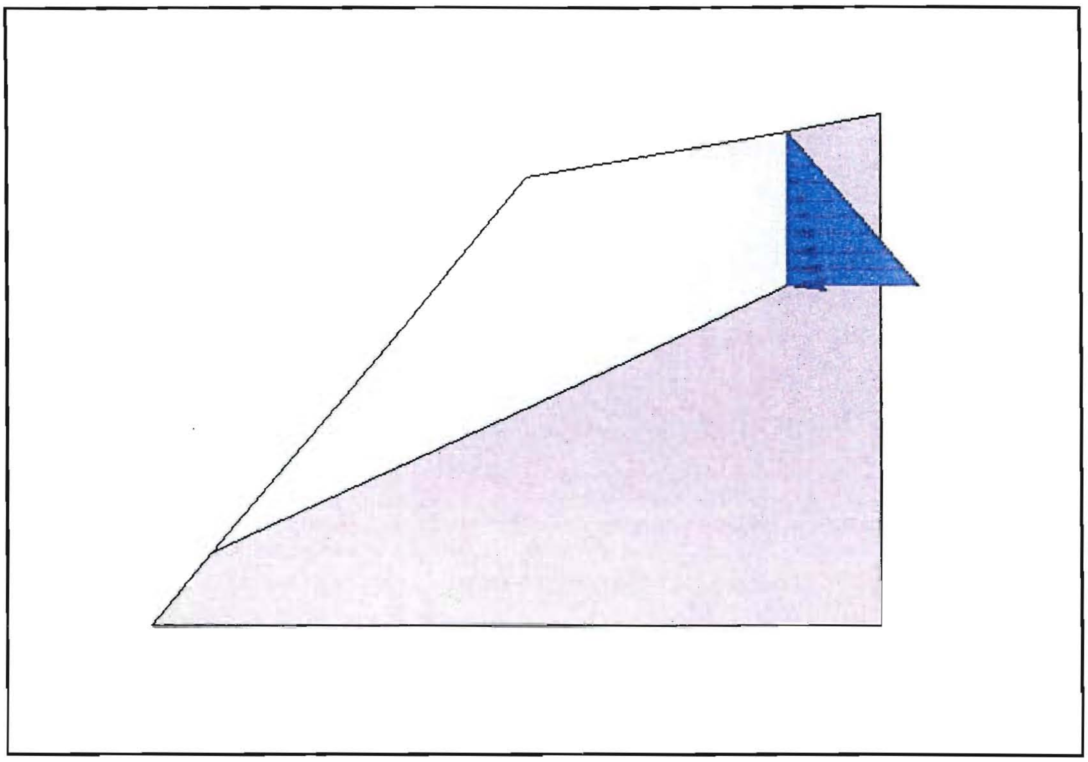
Hydrological models for slope stability



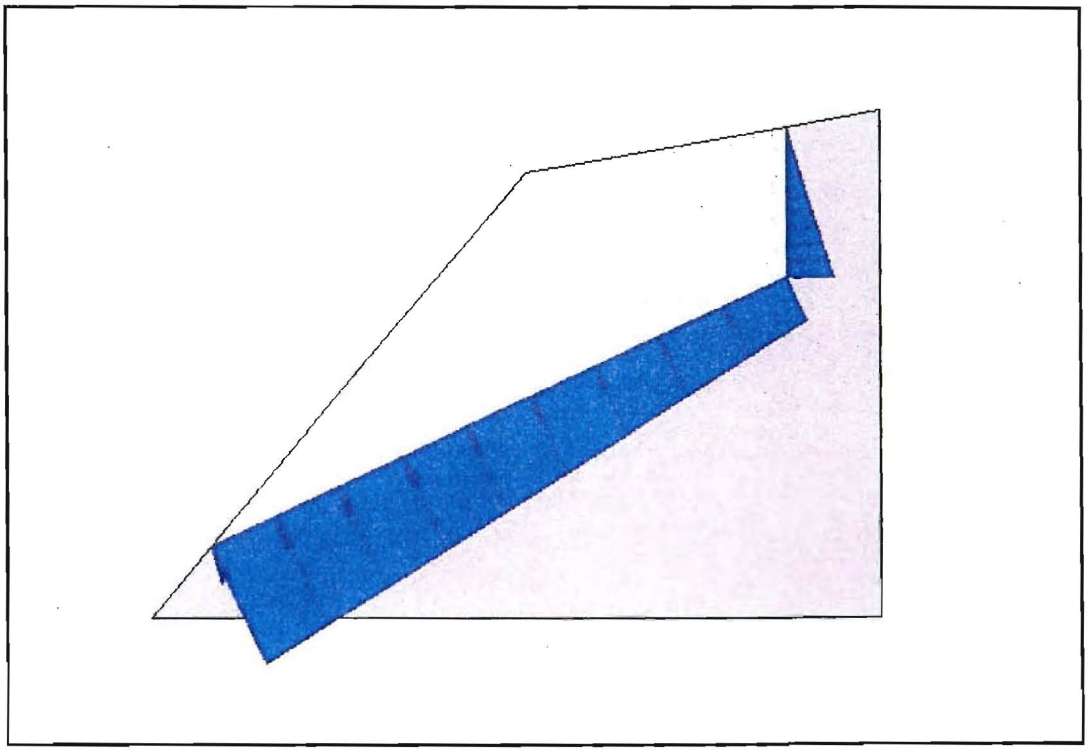
Peak pressure at mid height on the failure surface.



Peak pressure at the base of the tension crack.



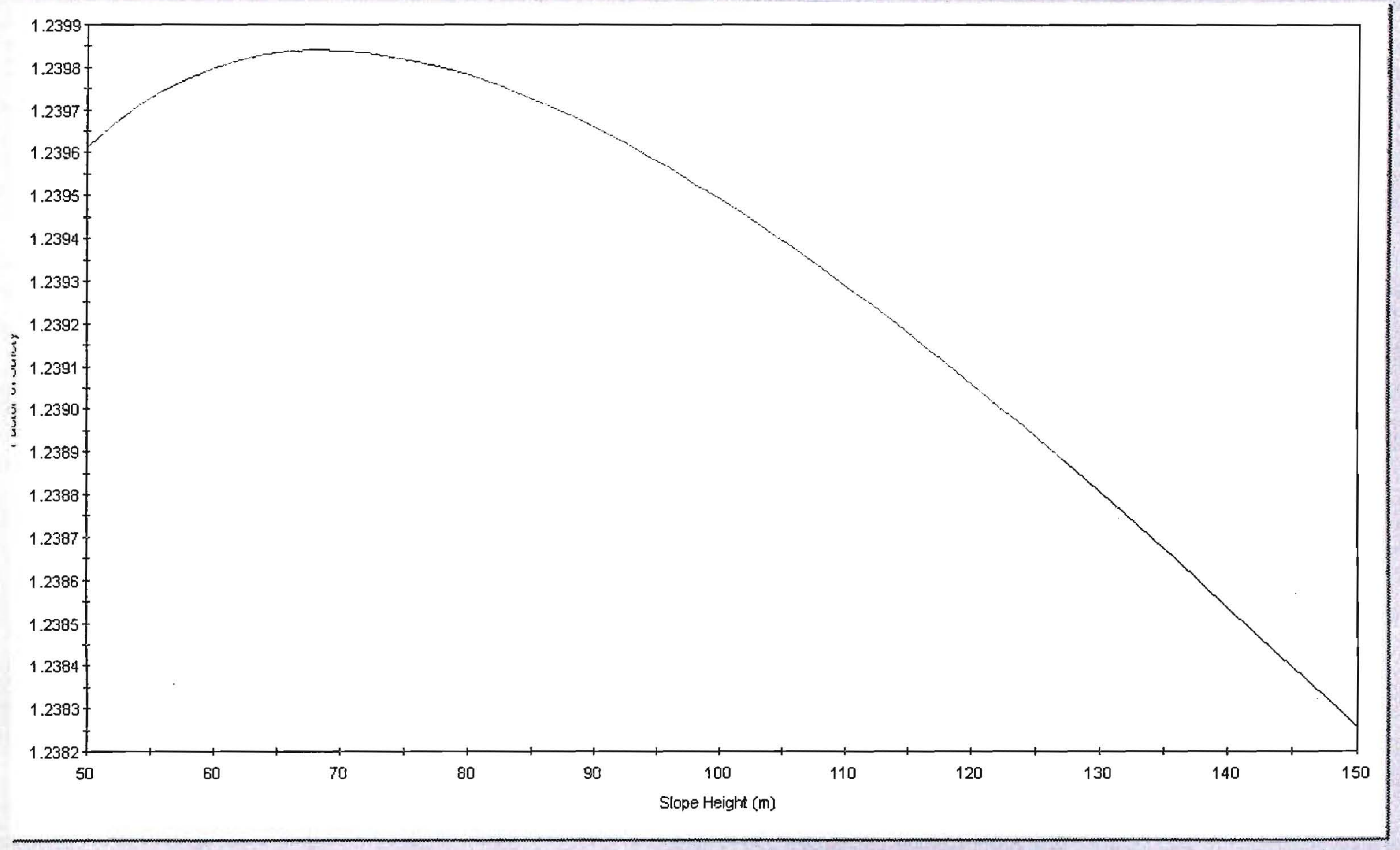
Peak pressure at the base of the tension crack with no failure plane pressure.



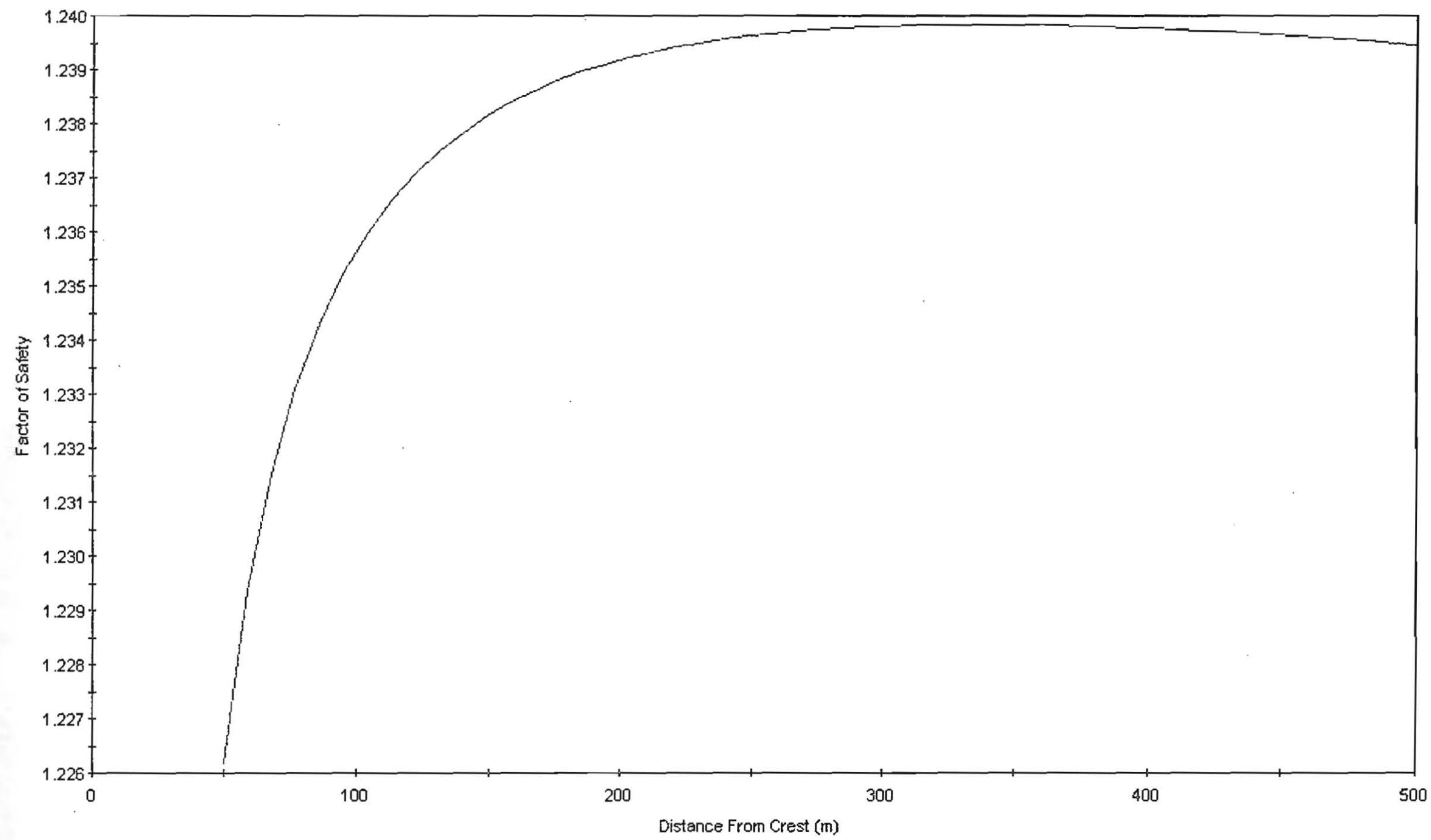
Peak pressure at the toe of the slide mass

Appendix III

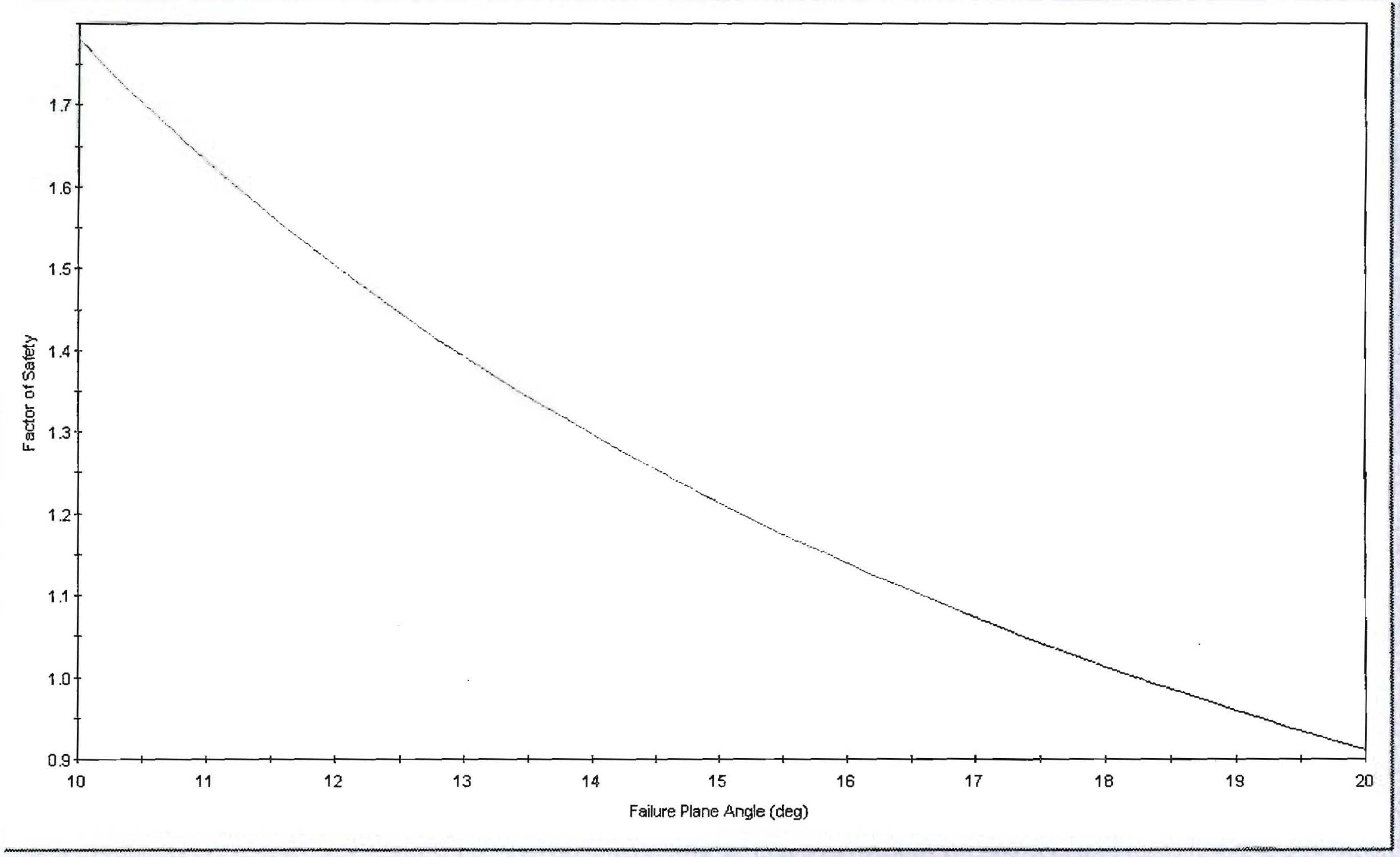
Ella Landslide sensitivity plots



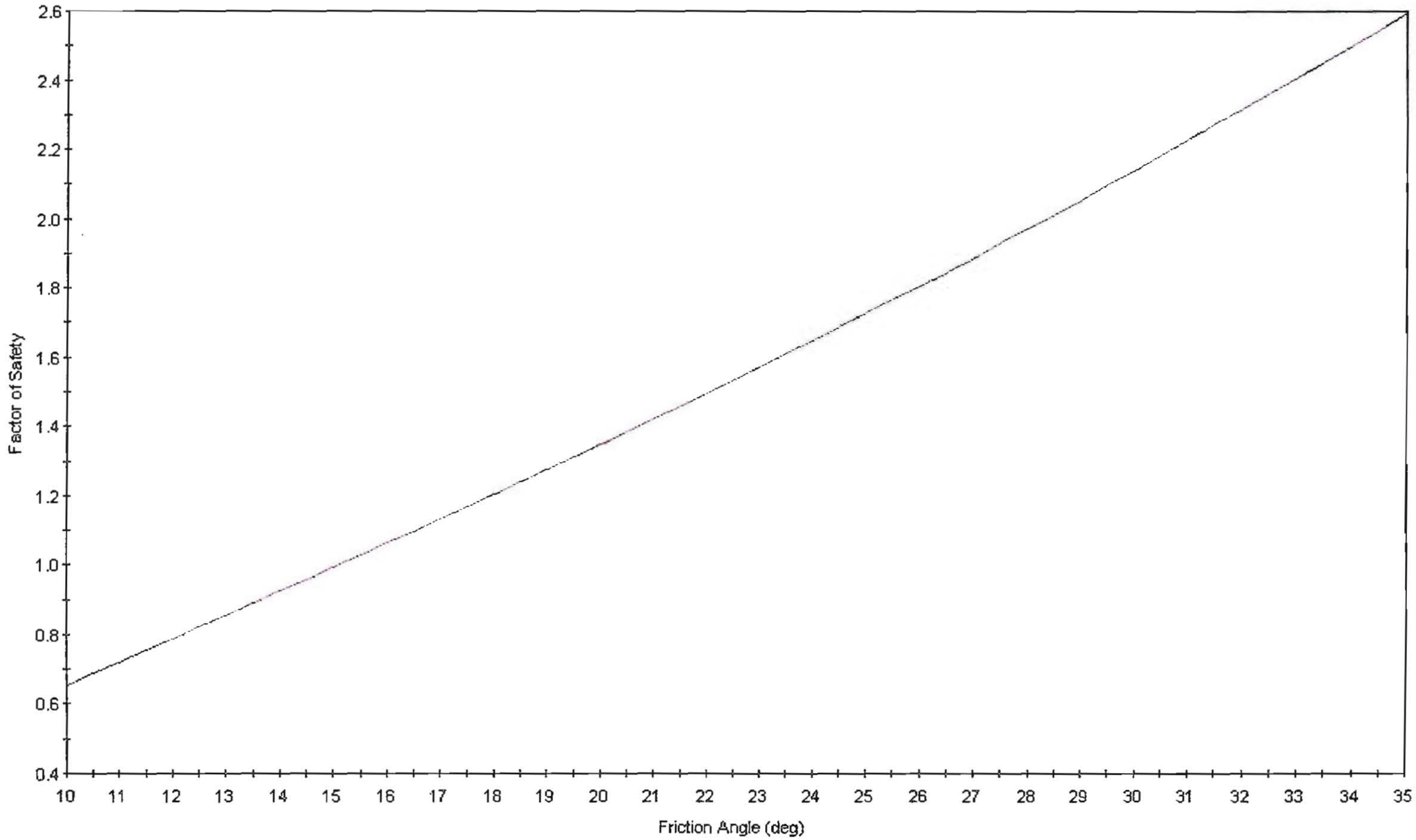
1.2399
1.2398
1.2397
1.2396
1.2395
1.2394
1.2393
1.2392
1.2391
1.2390
1.2389
1.2388
1.2387
1.2386
1.2385
1.2384
1.2383
1.2382
50
60
70
80
90
100
110
120
130
140
150
Slope Height (m)



Factor of Safety
1.240
1.239
1.238
1.237
1.236
1.235
1.234
1.233
1.232
1.231
1.230
1.229
1.228
1.227
1.226
0
100
200
300
400
500
Distance From Crest (m)

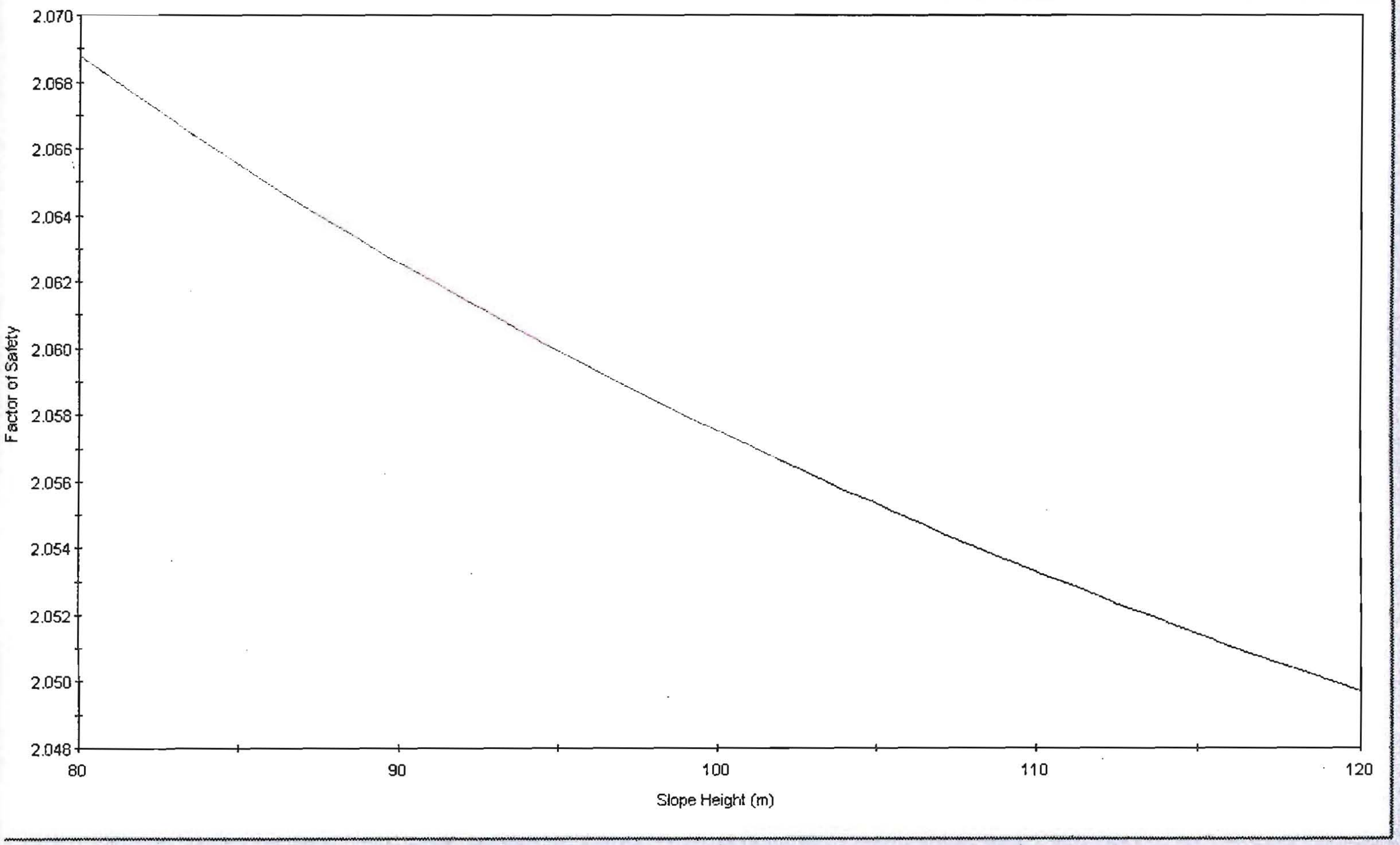


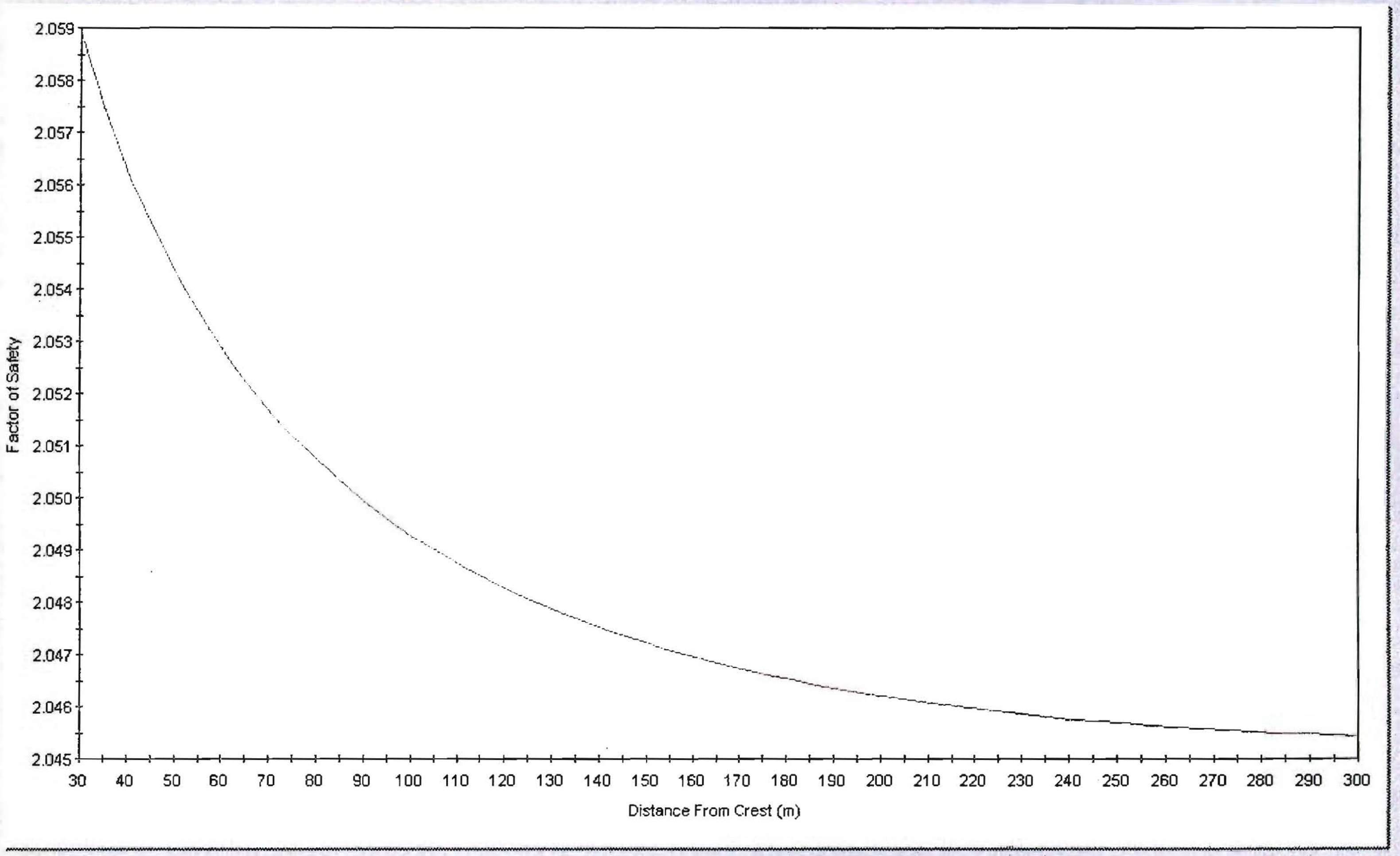
Factor of Safety
1.7
1.6
1.5
1.4
1.3
1.2
1.1
1.0
0.9
10
11
12
13
14
15
16
17
18
19
20
Failure Plane Angle (deg)


Factor of Safety
2.6
2.4
2.2
2.0
1.8
1.6
1.4
1.2
1.0
0.8
0.6
0.4
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
Friction Angle (deg)

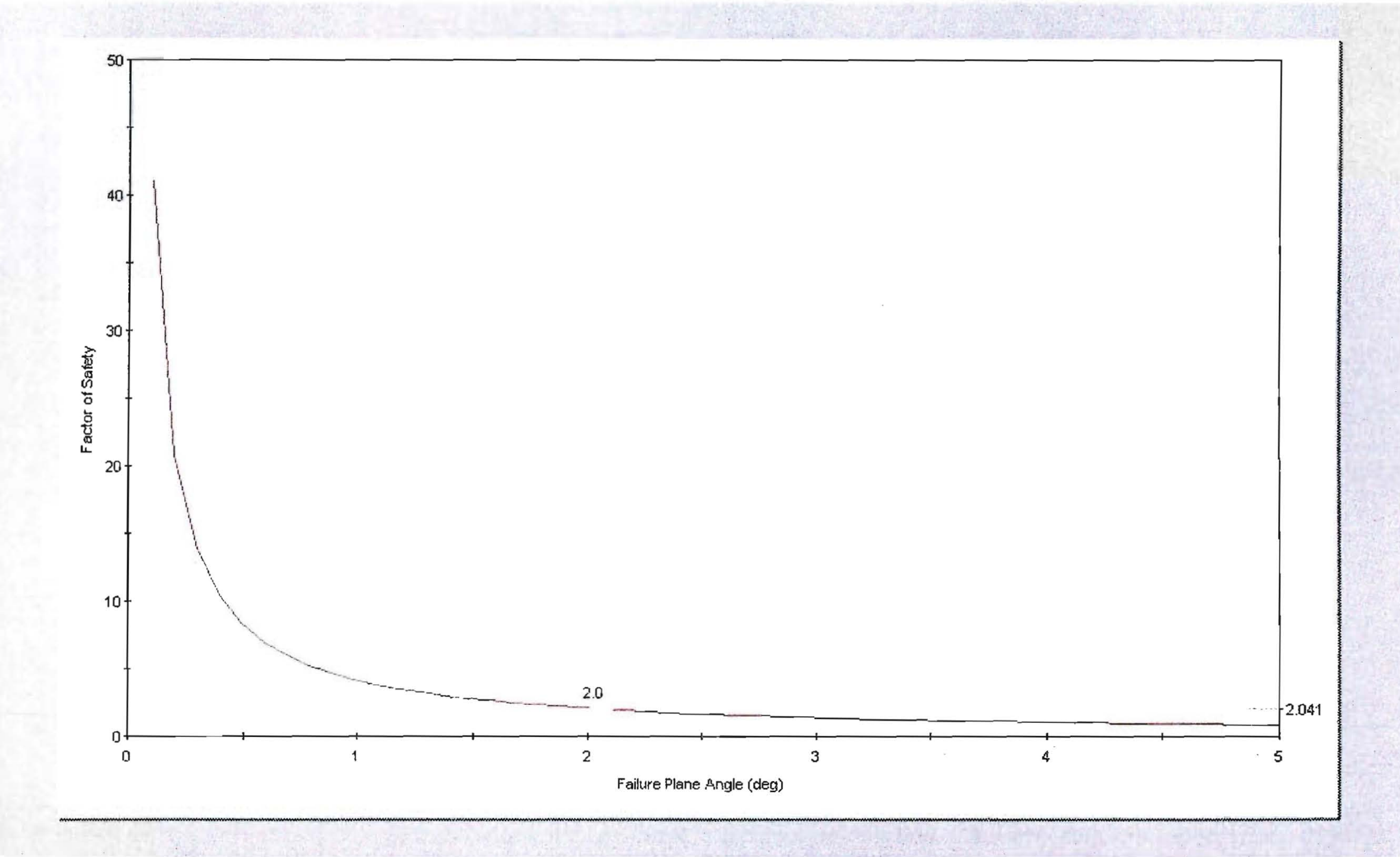
Appendix IV

Amphitheatre landslide sensitivity plots and probabilistic distribution

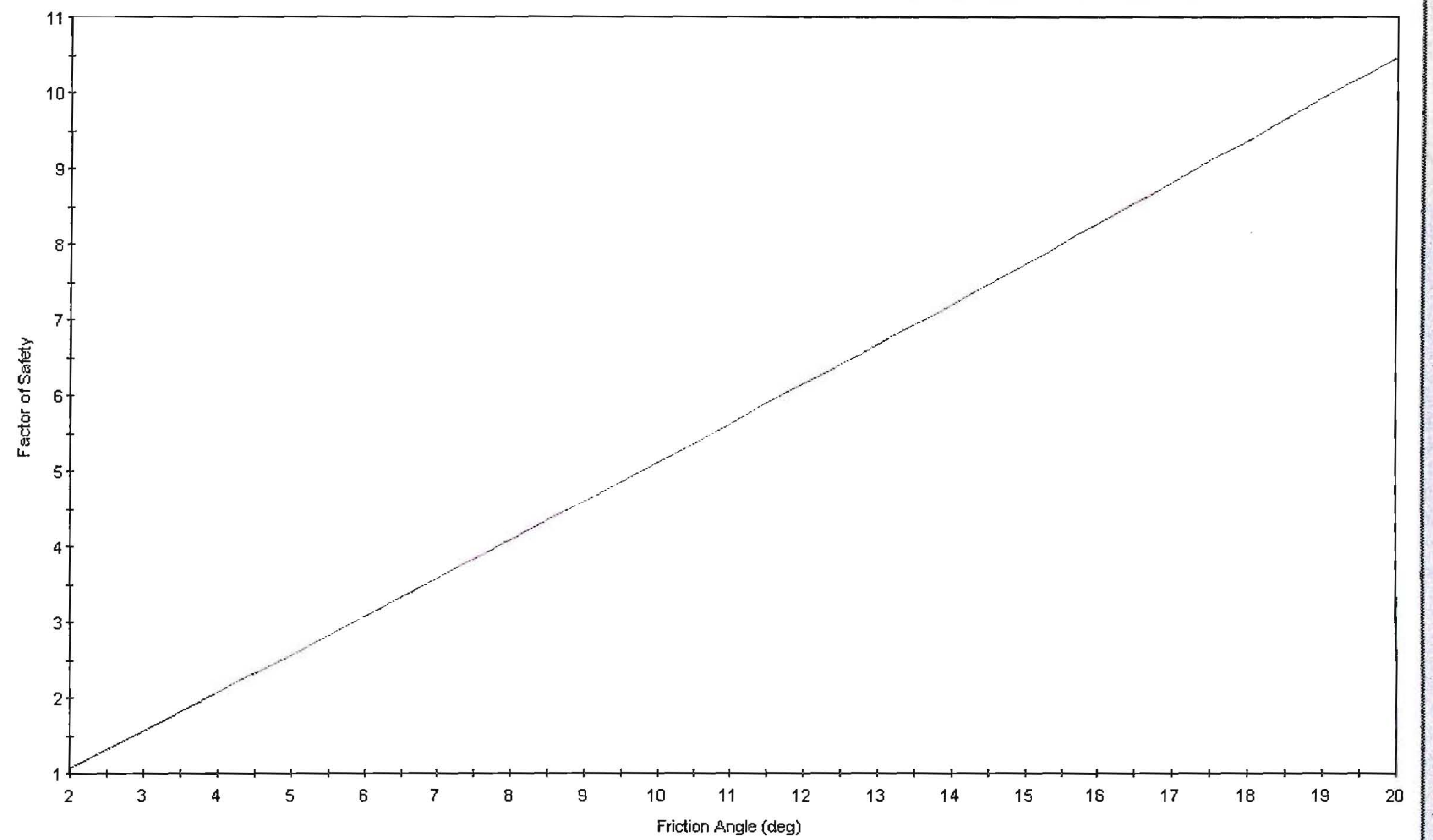

2.070
2.068
2.066
2.064
2.062
2.060
2.058
2.056
2.054
2.052
2.050
2.048
Factor of Safety
80
90
100
110
120
Slope Height (m)

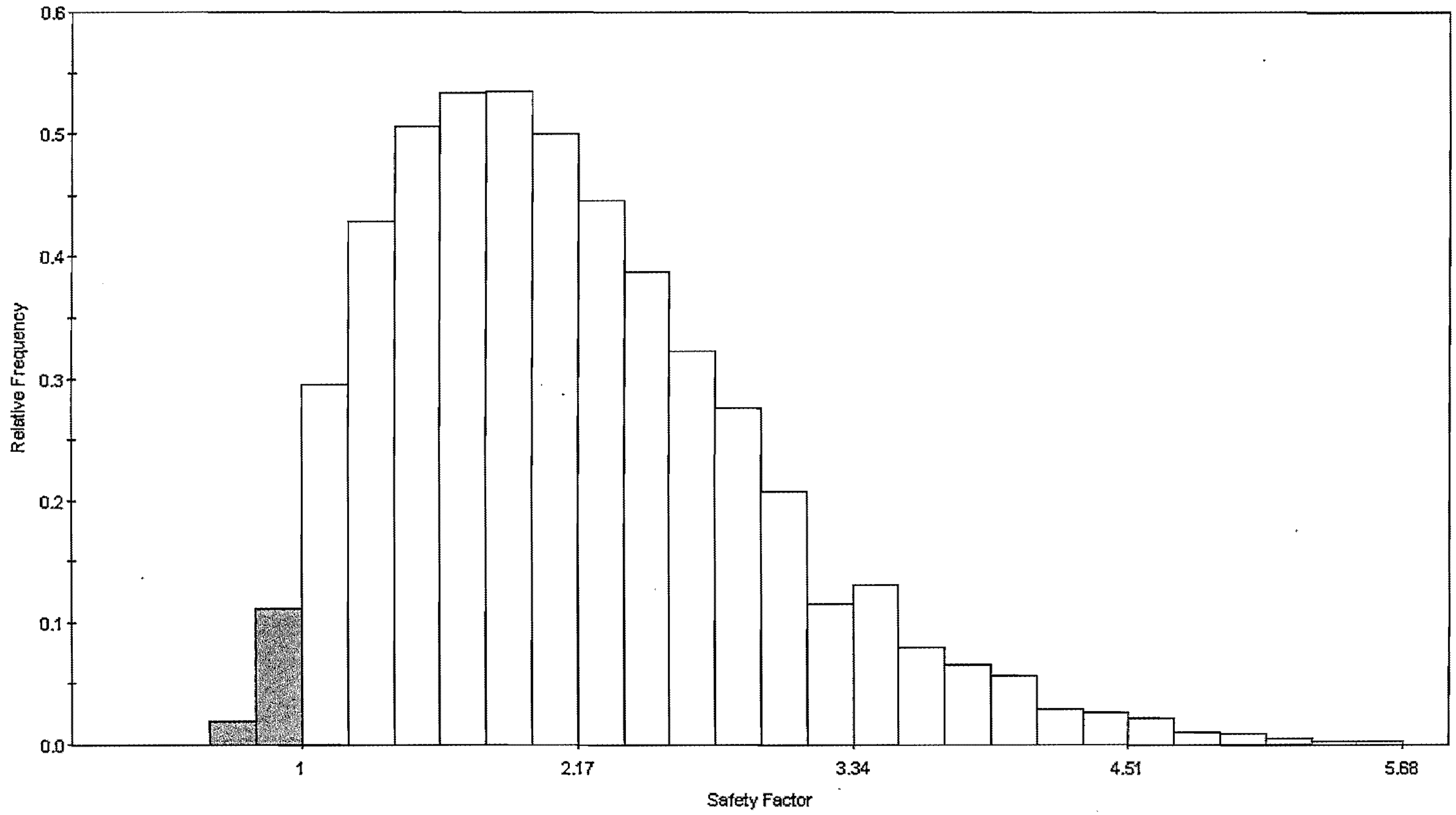


Factor of Safety
2.059
2.058
2.057
2.056
2.055
2.054
2.053
2.052
2.051
2.050
2.049
2.048
2.047
2.046
2.045
30 40 50 60 70 80 90 100 110 120 130 140 150 160 170 180 190 200 210 220 230 240 250 260 270 280 290 300
Distance From Crest (m)



50
40
30
20
10
0
Factor of Safety
0
1
2
3
4
5
Failure Plane Angle (deg)
2.0
2.041


Factor of Safety
1
2
3
4
5
6
7
8
9
10
11
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
Friction Angle (deg)



SAMPLED: mean=2.148 s.d.=0.7906 min=0.7052 max=5.578 PF=2.56%